VOLCANOES IN THE MOJAVE

David M. Miller, editor

2022 Desert Symposium Field Guide and Proceedings
April 2022
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Discoveries and adventures on the 1991 MDQRC field trip.
Volcanoes in the Mojave

David M. Miller, editor

2022 Desert Symposium Field Guide and Proceedings
April 2022
Captions

**Front cover:** Eruption at Fimmvörðuháls at dusk, photo by Henrik Thorburn, 27 March 2010.

**Back cover (top):** Pisgah lava field variations: Scoria cone in distance, pāhoehoe flow studded by white Ambrosia shrubs in middle ground, and 'a`ā flow in foreground, photo by D.M. Miller 2021.

**Back cover (bottom):** Interior cone rims within Amboy Crater, photo by R. E. Reynolds, 1993.

**Title page:** Cauliflower eruption through the roof of a Pisgah lava flow, photo by D.M. Miller 2021.

The Desert Symposium mourns the loss of
Ernie Anderson • Robert Bennett • Robert Hilburn

We also wish to note the loss of prominent desert researchers:
Douglas Morton • Michael Rosen

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The Desert Symposium is a gathering of scientists and lay people interested in the natural and cultural history of arid lands. The meeting comprises scientific presentations followed by a field trip. The Desert Symposium and its field trip take place annually, usually in April. The Desert Symposium publishes a volume of papers and a field trip road log. Safety, courtesy, desert awareness and self-reliance are expected of all participants.
A color version of this and past volumes may be viewed at http://www.desertsymposium.org/Publications.html
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Memories of Robert E. (Bob) Reynolds

Bruce W. Bridenbecker
Copper Mountain College, Joshua Tree, California

Introduction
As I began the journey of remembering Bob, it became apparent that he touched the lives of many. I remember the distant look in his eyes when he would make comments about going out to the desert. Every time I went to see him, Goggle Earth was on the computer screen where he was viewing a desert landscape.

I would like to begin with the information we posted on the Desert Symposium Inc. website.

Robert E. (Bob) Reynolds: 1943–2020
The Desert Symposium is saddened to report the death of Robert E. (Bob) Reynolds. For the last five years, Bob had fought cancer with his wife Jennifer at his side. He died peacefully July 21 at his home in Redlands, California. Along with Jennifer, he is survived by his son Jedediah, daughter Katura, and four grandchildren. Bob had a long career as an acclaimed vertebrate paleontologist and mineralogist, and was an extraordinary teacher and field trip leader in the wilds of the Mojave and Colorado Deserts. He knew the desert better than most, and he will be sorely missed by all that relied on his expertise, not to mention his boundless enthusiasm for the desert and its special places. Bob’s infectious good nature and love of the desert led many of us to a deeper understanding and love of the desert.

A mineral, an extinct snail, and an extinct gopher were named in honor of his contributions to mineralogy and paleontology. The mineral is called reynoldsite, \( \text{Pb}_2\text{Mn}^{4+}\text{O}_5\text{(CrO}_4\text{)} \) (Kampf et al., 2012) and occurs at the Blue Bell claims near Baker, San Bernardino County, California, U.S.A. and at the Red Lead mine, Dundas, Tasmania, Australia. \textit{Vallonia reynoldsi} (Roth and Reynolds, 1989) is a snail (mollusk) that was collected at Kokoweef Cave in the Ivanpah Mountains, San Bernardino County, California. \textit{Reynoldsomys timoteoensis} (Albright, 1997) is a gopher that was collected in the San Timoteo Badlands, Riverside and San Bernardino counties, California.

Bob’s service to students and researchers was honored by the Desert Symposium Inc. by the establishment of the Robert E. Reynolds Student Research Award. It is awarded annually to students involved in desert research.

Professional affiliations and awards
• President, past-president, and director, Southern California Chapter, Friends of Mineralogy National Board of Directors, Friends of Mineralogy
Memories

Memories and anecdotes were shared by Bob’s colleagues and friends, including Catherine Badgley, Sue Beard, Kristin Berry, James Bowden, Fred E. Budinger, Jr, Nate Carpenter, Steve and Val Castor, Margery Coombs, Jim Cornett, Mona Daniels, Stephanie Dudash, Walter Feller, Lowell F. Ford, John Foster, Dennis Jennings, Bruce Lander, Katharine Loughney, Spencer G. Lucas, Mike McKibben, Norman Meek, Chris Menges, Jeff Myers, Michael Patchen, Geof Phelps, Harry M. Quinn, Jennifer,

The Last Homecooked Rattlesnake: On a recent visit to the Reynolds’ house, I asked Jennifer and Jedediah to share a favorite Bob story. Bob began working as a mining geologist after graduating from the University of California, Riverside and his job included wrangling rattlesnakes from the entrance to mines in which he was trying to take samples. Thus, he knew what to do when he was surveying in the San Bernardino Mountains, opened the door of his vehicle, began to put his foot down, and saw a rattlesnake right where he wanted to step. He was able to get the snake into a burlap bag which he then repeatedly hit with a shovel to ensure its demise. When he left, Bob placed the snake in an ice chest, brought it home, and placed it in the freezer. The next day he brought out the frozen solid rattlesnake and asked Jennifer to cook it for dinner. He put it in the kitchen sink to thaw out. When Jennifer was ready to prepare the snake for dinner, it was not in the sink. A rather frantic search found the snake behind the refrigerator, very much alive. Needless to say, it was the last time homecooked rattlesnake was on the menu.

I first met Bob out by the fireplace dishing up his homemade chili on Thursday night. Being new to the area, I did not want to be far behind Bob when we started off through the desert. I soon learned that even though I was the first car behind him, there was no way for me to keep up. All I could see was a cloud of dust and hoped that I didn’t lose that cloud. One often did not know what was around the bend. Many will remember the year of the flat tires, when we showed up in Brawley with many vehicles either needing a complete new set of tires or at least some tires repaired when Bob led us down a road full of old railroad nails. The tire store owner opened the store early and we probably gave him a good portion of his income for the year.

A Bob quote. It is okay to get in over your head, just be prepared for what may happen.

Bob Reynolds was a long-time friend. I miss him greatly. Bob graduated from UC Riverside just before I got there in 1966. As I began my work on Tertiary mammals of the Mojave Desert, I became aware of Bob’s contributions and information and we started sharing those things on numerous times together, especially on trips of the Barstow Formation and other units in the Mojave. Not only did he understand the geology, stratigraphy and vertebrate paleontology of the Mojave Desert, but Bob was a wonderful, observationist and humorist of life in general, including his later times thereof. Being with Bob was always fun, in the best senses of the word. I know I am not alone in having this experience and these feelings. The memory of Bob Reynolds shall long persist.
For many years Bob Reynolds urged me to write up the chalicothere material he collected from the Cajon Valley Formation. In 2015 the topic for the Desert Symposium was just right to move me to collaborate with him on this project. I’m grateful that I had the opportunity to publish this paper with him. Even more I had a great time attending the symposium at Zzyzx and going on the associated field trip, even if I bounced around incredibly for miles in the back seat of his off-road vehicle. It was great knowing you, Bob. RIP.

I was fortunate enough to spend a day in the field with Bob, Ev, and Jerry after the last Reno SVP meeting. I remember Robert was very happy because he just got a new field vehicle. We spent the day on a personal tour of the Muddy Creek Formation sediments. Robert was always generous with his knowledge, time, and publications. I miss him already.

A Bob quote, watching friends set up camp with folding chairs, tables, and cots: The whole desert is my armchair.

I was incredibly saddened to hear of Bob’s passing. And selfishly even more so that I hadn’t met Bob earlier in my life and had the opportunity to spend more time with him, going out on fossil hunts and eating his chicken soup. As many have pointed out, Bob had an incredible wealth of knowledge regarding Paleontology and minerals. I once took boxes of minerals I have over to his house for help with ID’ing them and he was able to not only ID every one of them but also tell me which mine they each came from. The last email I received from Bob was “Yes - I’ll never stop enjoying the desert”. As surely as Bob will be missed, as long as I live, he will live on with me wandering around the desert.

I remember him as an extremely knowledgeable person who was always willing to help people out. The profession has lost a valuable resource.

He was a marvelous person, so ‘field-fun’ and ‘field-enlightening’. I learned so much from him. A truly wonderful human being—one of my first field stints with him he introduced me to post-Dire Straits Mark Knopfler. And I was so in awe that he was friends with Mary McCaslin; I was listening heavily to her at the time. He will be deeply missed.

I am greatly saddened to hear that Bob has passed on. He was a great friend and inspiration. I will sincerely miss our chats and sharing of desert experiences. Above all, his extremely informative and carefully planned road trips, each an adventure all in their own.

Bob, to me, was an incredible resource. I was fortunate to work and publish with him in the mid-1990s. Paleontology has lost one of its great students with his passing.
Bob was a great friend and colleague and life was never more interesting than when talking to Bob. His commitments and dedication to the science of paleontology as well as to the faculty, scientists and students as well as the public was boundless and I will miss him very much.

Bob is going to be a great loss to the geologic community, particularly to those of us with a love of the Mojave Desert. I considered him both a personal friend and a geologic colleague. You probably know that Bob, George Jefferson and I all go back to the late 1950's to early 1960's at UCR. We spent many a weekend collecting vertebrate fossils in the desert when we probably should have been in class studying. Bob will be sorely missed.

I am so glad that we got to spend a typical Bob-day with him last April (2019) driving to remote outcrops of what he thinks is Barstow Formation. What a character. He made such a difference in helping us get started with field work in the Barstow and Cajon Pass areas.

I remember him as a kind and knowledgeable individual.

The Desert has lost a giant! Zzyzx has lost the best field trip host ever and we’ve all lost a great friend.

We lost a collective knowledge of the desert that few of us can comprehend. He will be sorely missed. A good man, may the love for the desert live on.

I know that myself and many of us here at the Needles Field Office will be upset with the passing of our Dear Friend. He was an inspiration to us all, and gave us the excitement of new discoveries each time we saw him. He was a wonderful man.

Bob will be greatly missed. He was a warm and inclusive person. He was always making sure that students at local educational institutions had multiple opportunities to attend symposia and field trips.

Bob possessed a tremendous wealth of knowledge and always delighted to share and inspire. And he was one heck of a fine writer and editor.

Thank you for the kind words for this extraordinary man.

For decades, Bob hosted Thanksgiving dinner for friends, family, and colleagues at his favorite campsite in the New York Mountains of the eastern Mojave Desert. He’d build a fire and let it burn down to coals so he could roast the turkey. This tradition always included “modifying” the turkey with a sledgehammer so it would fit into the Dutch oven.

Bob will always remain in my mind an icon of the Desert Symposium and Mojave fieldtrips, with his booming voice.
and big, welcoming personality, and as someone observant and full of good ideas. We are truly better for having him as part of our Mojave life. He’s still with us.

I last saw Bob and Jennifer at their home in Redlands when Bob had a barbeque for friends from years gone by, many of whom I hadn’t seen since 1985, when I worked for Bob at the museum.

I knew Bob for more than 40 years and I always considered him a friend, advisor and mentor. For more times than I can remember, he kept me (a non-geologist) and my museum on the right track when we were developing exhibitions and printed materials involving geological issues and concepts.

My fondest memory of Bob was when he, myself and author David Darlington spent three days traversing the Mojave Road for a chapter in David’s book The Mojave. On the first evening, Bob prepared dinner by scooping out a hole in the sand, tossing in a raw pork roast, covering the meat with wood and lighting the pile on fire. “But Bob, the meat will be covered with sand and soot,” I exclaimed. “It will be fine,” Bob replied. After about an hour (as I recall), Bob cleared away the burning embers, brushed off the sand and ash and served up the best pork dish that I have ever had.

If it were not for Bob, my research pursuits would have waned over time. His constant persistence with the Desert Symposium and encouraging me to submit papers kept my eye on the ball and kept me gathering field data. Any information I have revealed about the desert is, at least in part, due to Bob.

Our presence on Earth is so very brief. But sometimes a person comes along whose existence transcends their biological life and whose influence persists long after their passing. Bob was such a person. The desert and all those persons who have passion for it, have a bit of Bob within. He will be missed but never forgotten.

I am very sad to hear about the passage of Bob Reynolds. He was a wonderfully engaging and energetic person with a vast, long-lived, and nearly unsurpassed knowledge of the geology, natural history, and culture of the entire Mojave and lower Colorado region that he so willingly and enthusiastically shared with so many others of all ages and expertise. One of his major, but not only, contributions was his unique persistent sponsorship for many decades of the Desert Symposium, which I had the privilege of attending and contributing to a number of times starting several decades ago. I always enjoyed the unique combination of collegial interaction and interpersonal comradery that characterized all of the symposia that I attended. I will always fondly remember and treasure my time with him.

Bob Reynolds was a singular presence in the world of Mojave Desert geology and paleontology. I first met Bob in 1971 when he was the young curator of geology at the San
Bernardino County Museum. This was when the museum was in Bloomington. In 1974, I was working at the Calico Early Man Site. Bob and everyone else at SBCM were busy moving every rock, fossil, and artifact from Bloomington to the newly constructed museum in Redlands. I helped, as I could, providing many pickup truck loads of desert rocks and sand for various archaeology displays being set up by Ruth DeEtte ("Dee") Simpson, the curator of archaeology. Bob and his volunteers were busy with geology and paleontology displays. In those days, archaeology and paleontology occupied adjacent areas in the basement of the main building at SBCM; I got to know Bob fairly well.

Bob was very knowledgeable and always willing to share. He exhibited that throughout his life.

Bob and his crews conducted many paleontology projects all over the Mojave Desert. A couple of times they stayed at the Calico Site. I clearly recall the time they were working on Solar One, the county’s first big solar plant, at Daggett. Bob knew how to live well in the field. He was a good cook; he knew how to eat well, and he always shared the good times.

Bob Reynolds took the reins of the fledgling discussion group that we now call the Desert Symposium in 1987. My friend, Bob Adams, and I organized the first event as a "beer & pretzel bull session" regarding Mojave Desert investigations in Reseda in August 1985. The next year, George Jefferson hosted the event at the Page Museum at the La Brea Tar Pits in Los Angeles. The third year, Bob Reynolds took the reins and the event came out to SBCM in Redlands. In those days we called it the MDQRC Symposium. This fancy title stood for "Mojave Desert Quaternary Research Center" symposium. After Bob left SBCM, he and Bill Presch from Cal State Fullerton brought the Desert Symposium out to Soda Springs (Zzyzx). The rest is history. We can all treasure our own unique ensemble of memories of Bob, each in our own way.

Once, a week before spring break, Jenny received a phone call from the kids’ elementary school principal asking if Bob was okay. He was concerned because the school had become accustomed to Jed and Kate being “sick” before a scheduled school vacation – Bob liked to beat the traffic and maximize the time available for family camping.

Many years ago, in the 1990s I was leading a Quaternary seminar for a small group of CSUSB students. We were in a very remote area between Mountain Pass and Cima, and decided to pull over for the night along a barely usable dirt road near a mountain base. The clear night got quite cold and when the sun came up we stayed in our sleeping bags waiting for it to warm up a bit. While we were doing that, to my surprise an old truck slowly came down the road. I recognized the driver as Bob Reynolds, and he stopped and walked over to visit when I waved. I asked him why he would be up so early driving through this very remote area and he said that he was looking for dinosaur tracks. Just then he pointed behind me and casually mentioned there


Kate, Jed, and Jenny Reynolds carrying empty dynamite boxes from the Mohawk mine, Clark Mountains, 1979.
was a very large rattlesnake on the large rock surface not far behind me. Thinking he was just joking, and without looking, I responded that it was much too cold for snakes to be out. He then told me to turn around slowly. Less than 20 feet away from where I was standing (and where I had been sleeping), there was a very large rattlesnake resting in the center of a large sunlit rock surface warming up. Needless to say, my class was shocked by the surprise development. Thanks for teaching me (and my class) about snakes, Bob!

There are few people in this world with such deep passions for their work and Bob was one of them. He inspired me and will always have an influence in my life in many ways. He gave me opportunities that I will always be grateful for. I tell people that Bob knew the desert like the "back of this hand" and was always willing to share in its secrets. I still catch myself thinking "I will need to tell Bob about this…” or "I wonder if Bob could tell me more about this …” He will be missed by many. Bob’s passing marked the end of an Era. He passionately believed in geology and paleontology field work, lab work, collections, and sharing the knowledge of what he learned from all of it. His passions live on in all of the many lives that he touched. Mine is definitely one of those!

Bob will be missed. We’ve known him since the early 60s. He was always so jovial and so enthusiastic about his science and hobbies. He was uniquely both macho and kind; also non-judgmental—I don’t recall him putting people down, professionally or otherwise. Bob was magnetic and it seemed he always had a crowd of people around. I recall that his field camps were particularly popular, although he also seemed happy alone in the desert. I was amazed at Bob’s ability to identify vertebrate fossils using a single fragment, and think he was a real genius at that.

According to Bob’s mother, family vacations were inevitably rerouted because an interesting mine or a fossil locality was “nearby” (often in an adjoining state). As soon as Bob got his driver’s license, he could be counted on to be late for appointments because on his way he’d always have to stop at road cuts to check out the geology or collect a specimen or two. If anyone stopped to see if he needed help when he was parked on the side of the road in a remote place, they’d be surprised with an impromptu lecture on the local geology.

**Three Mojave girls salute Bob Reynolds**

Catherine Badgley, University of Michigan
Tara Smiley, Stony Brook University
Katharine Loughney, University of Georgia

**Catherine:** In 2012, when I decided to begin a field program in the Miocene sequences of the Mojave region, it was Bob who jump-started the realization of plans for me and my students. Our goal was to study the effects of the middle Miocene warming interval from 17 to 14 million...
years ago on the diversity and composition of mammalian faunas.

Initially, the Barstow Formation was our focus. Bob introduced us to the Crowder and Cajon Valley formations, which also spanned the middle Miocene. His introduction to the formations in Cajon Pass and the Barstow area consisted of an extended tour—in his trusty pickup and on foot—of access routes, outcrops of the formations, and fossil localities, as well as meetings with people in charge of permits, permissions, or potential accommodations during our field seasons. It would have taken us several years to make all of these connections on our own. With Bob as a facilitator, we were able to get to work after one year of planning and successful application for field permits. Tara Smiley conducted two field seasons in the Crowder and Cajon Valley formations as the basis for her PhD thesis, and Katie Loughney conducted three field seasons in the Barstow Formation for hers.

During each of our field seasons through 2019, Bob would spend a day or two with us—showing us ever more remote and far-flung outcrops and fossil localities that needed investigation. He also invited us to the Desert Symposium and made our research interests the central theme for the 2015 symposium (Mojave Miocene) and again in 2020 (Changing Facies). A fond memory is of Bob with his beloved pickup, full of camping gear, vegetable trays from his cooler, a portable office with maps, papers, and field guides, the extra spare tire, and field gear.

**Tara:** From my first meeting with Bob Reynolds, he has been a vast source of knowledge about the paleontology and geology of Southern California and an enthusiastic and tireless proponent of sharing and building upon that knowledge. Bob played a critical role in the development of my dissertation research at the University of Michigan. In 2012, he introduced me to my field area in Cajon Pass, where the fossils he had meticulously collected, along with associated stratigraphic and geological information, became the basis for much of my dissertation work. In the years that followed, he always joined us for a day or two as we worked in the region. On those lucky days, Bob’s itineraries for us were truly impressive. We’d traverse many miles, drainages, dusty powerline roads, and overlooks to see the outcrops and field sites that Bob knew like the back of his hand. At breakneck speed, Bob would share decades of observations from the field, while my attempts to keep up with him led to pages of text in my field notebook, the words hoping to capture his deep knowledge of the landscape and its history, the exclamation points around certain topics capturing his excitement over unanswered questions, and near indecipherable scribbles reflecting the bumps in the ‘road’ as he zoomed from one remote outcrop to another. It was always an adventure.

As a graduate student and beyond, I was not alone in receiving Bob’s incredible support and encouragement – he has influenced a long line of students before me, and I am happy to build on that body of work generated in large part due to Bob. In addition to his paleontological and field
knowledge, Bob also shared with us a fantastic network of collaborators and local contacts. He facilitated fieldtrips with some of my paleontology heroes, helped us establish important contacts with BLM officers, and connected us with friendly faces to help in screen-washing, lodging, and transportation to remote desert localities. Among the many individuals Bob introduced us to, all expressed much respect and appreciation for Bob and his work, evidence of his far-reaching impacts. Beyond this generosity of time and knowledge, Bob was incredibly kind and supportive, always treated me as a colleague – even as an early graduate student – and he valued and championed contributions from myself and others in the group. I am truly honored to express my appreciation for Bob and thank him for his passion, efforts, and contributions to the field of paleontology and his mentorship of myself and countless other individuals.

Katie: I first met Bob Reynolds in the summer of 2013 when I joined Catherine and Tara in the field as a new graduate student. He drove up to meet our field group at Owl Canyon Campground for lunch, where he answered our questions about anything and everything Mojave-related while deconstructing a rotisserie chicken. Afterward, Bob drove us to several locations in the central Mud Hills, where he seemed to know every canyon and gully. It was my first introduction to what would become my project field area, and although this area was very familiar to Bob, his interest and curiosity were as fresh as mine. His well-used and well-stocked pickup truck spoke volumes to his experience and comfort in the field. One of our field assistants that summer was so taken with Bob that he bought a blue cotton button-down shirt to wear in the field. That meeting marked the start of an annual tradition of day-long field trips with Bob, which regularly covered many miles and several million years of Mojave history.

Over several field seasons, Bob introduced us to the geologic history, stratigraphy, and paleontology of the Mojave and also showed us the best exposures and how to access them. My field notebooks quickly filled with copious notes and maps shakily drawn as Bob sped down sandy BLM routes in his pickup. He would slow down occasionally to note the mileage or to advise, "Don't forget to think," pointing to an enigmatic sign at a road junction. His guidance was instrumental in helping me begin my research in the Barstow Formation, and had it not been for Bob, my dissertation work would have taken me longer to complete and would also have been less comprehensive. One field season, Bob drove us up to a difficult-to-access part of the Mud Hills. He pointed out one especially productive fossil locality, saying, "I spent my life there." I am lucky to have known Bob and to have benefitted from that lifetime of work.

All of us: Ten years, two PhDs, new collaborators, and a dozen papers later, we are still following in Bob’s actual and virtual footsteps.
I first met Bob at Pasadena City College while taking classes in geology. Besides excellent teachers we had the Dana Club, a loose knit but wonderfully active semi-organization. We tried and usually succeeded in going out every weekend, save finals week, to some mine or location of interest.

At one point Bob, Mike Mauer, and I were sitting in a pleasant place just outside of the Calico Mine, which was the centerpiece to Calico Ghost Town but had been essentially undeveloped at that time. We had just spent about three hours in the mine, entering through an unguarded entrance. The exploration had been strenuous, at one point going up the oldest and highest of all wood ladders that I have seen, on the face of a very large stope. The ladder was pinned to the slickenside surface of an old fault line and seemed to have no end or safe place to get off. We finally surfaced and sat at the pleasant place mentioned above, covered with the dirt and smudge we had acquired in our exploration and cleaning our carbide lamps. (This was at a time when electric mining lamps were terrible expensive and had a short-lived battery life. Justrite carbide lamps were both available and reliable, if prone to sludge, smudge, and soot.)

Then we saw a strange apparition in the path below us. It bobbed up and down as it made its way to us and eventually resolved to reveal a park ranger in full uniform. He said, “I hope you boys aren’t planning on going into the mine. It is closed and it is very dangerous.” We assured him that we had nothing like that in mind and he left cheerfully, having done his duty.

This was the only time that those of us in the Dana Club had any contact with officialdom, and it resolved itself. We were great friends and had a great education coming in large part from the Club itself.

Much more could be said. On behalf of the Desert Symposium Inc, I want to thank everyone for their comments. Some say he's out there still.

References
Volcanoes in the Mojave: the 2022 Desert Symposium field trip road log

D. C. Buesch, D. M. Miller, B. W. Bridenbecker, and M. R. Sweeney

Introduction
Basalt lava fields, some decorated with scoria “cinder” cones, are scattered around the Mojave Desert (Fig. I-1). Most basalt fields are short-lived, but the Cima volcanic field is unique in having eruptions that span ~7.5 m.y., including the youngest eruption in the Mojave Desert at ~12 ka. Xenolith-bearing basalts that include both mantle and deep crustal rocks are known in several fields. All basalt fields except Cima are restricted to the active eastern California shear zone, and many lie directly on active faults, indicating a direct relation between faults and volcanism (Miller and Buesch, 2022). Area and volume of lava is greatest for those fields associated with dextral faults, which may be attributed to less compressive stress across those faults and less resistance to buoyant rise of magma, as compared to sinistral faults. Several of the large fields lie along a broader WNW trend from Amboy to Pisgah. The WNW trend from Amboy to Pisgah fields is associated with topographic lows, possibly

Fig. I-1. Map of the young (<7.5 Ma) basalt fields in the Mojave (from Miller and Buesch, 2022). The Desert Symposium 2022 field trip will visit Pisgah, Dish Hill, and Amboy on Day 1, and Cima and Bicycle Lake on Day 2. The path of the trip is the black dashed line.
caused by right steps in the dextral faults (releasing bends) that would further enhance intrusion. Furthermore, small fields along east-striking sinistral faults may owe their diminished size to increased compressive stress across those faults, limiting intrusion by dikes. Deep shear zones associated with the faults may promote generation of basalt (Valentine et al., 2021).

Age patterns for the young basalts are complex but in the area of figure I-1, fields between 6 and 3 Ma span much of the area and Quaternary fields spread between Pipkin on the west and Amboy on the east. Miller and Buesch (2022) found that ages of young basalts fields in the Mojave Desert are clustered as four eruptive pulses spanning from ~9 to 6 Ma, from 5.4 to 4.7 Ma, from 3.9 to 3.1 Ma, and younger than ~2.1 Ma.

The Desert Symposium 2022 field trip will examine the geology of five basaltic volcanic fields in the central Mojave Desert, including the Pisgah, Dish Hill, and Amboy fields on Day 1, and the Cima and Bicycle Lake fields on Day 2 (Fig. 1-1). These fields are characterized by their youthful morphology, alkalic compositions, and in some, inclusions of mantle xenoliths. The youngest basalt we will visit is in the Cima field at 12 ± 2 ka (Phillips, 2003) and the oldest is the recently dated 4.55 ± 0.07 Ma Bicycle Lake basalt (Buesch et al., 2022). We will examine the physical volcanology of the basalt fields, including the types of eruptions (effusive and explosive) and the resulting deposits (lava flows, scoria cones, and tuff cones). These exposures provide excellent examples for how to interpret the more limited exposures of much older deposits. This field trip is not only a travel through the physical volcanology of basalt fields, but through the perspectives of petrology, paleomagnetism, remote sensing, and planetary geology. These volcanic fields have been used as classic examples in each of these types of studies, and in many ways, we will be walking through a museum. But first, a short review of physical volcanology.

Review of volcanic products and processes

Volcanic rock names
Field names for volcanic rocks, such as basalt, are based on the macroscopic (using a hand lens) identifiable minerals (phenocrysts) in an aphanitic groundmass (glassy or crystallized, no minerals identified with a hand lens). Volcanic rock names are also based on the whole-rock chemistry, and the first-order classification is based on ranges of silica and alkali contents (SiO2 and Na2O+K2O percent normalized to volatile-free weight percent, typically displayed on a Total Alkali Silica diagram e.g., Le Maitre, 1989) (Fig. I-2). The focus on this trip is on rocks with field names of basalt, but in detail this includes basalt, basaltic andesite, basanite, trachybasalt, and basaltic trachyandesite (Fig. 1-2).

What drives eruptions
Volcanic eruptions occur when magma from a deep source (typically envisioned as a chamber or an integrated network of dikes) moves upward along a fissure to the ground surface. As the magma rises the confining pressure on the magma decreases and gases that were dissolved in the magma begin to separate to form vesicles. The expansion in the number and size of vesicles increases the volume of the magma, and this in turn increases the upward driving force to form effusive eruptions of lava. If vesiculation is great enough, the vesicles interact and break, thereby fragmenting the magma into fragments of various size. Some fragments are ejected as liquid blobs that can develop aerodynamic shapes before they hit the ground. Explosive eruptions produce ash (<2 mm), lapilli (2-64 mm), and bombs or blocks (>64 mm); bombs are rounded, and blocks are angular.

Types of eruptions and deposits
Effusive eruptions have vents that are either long fissures or point sources and result in lava flows, whereas explosive eruptions have point sources resulting in cones of pyroclastic material. Valentine et al. (2021) reviewed large Quaternary basaltic volcanic fields in the American southwest, and although they did not include any of the fields from the Mojave Desert, the summary of types of eruptions and deposits applies to this region as well. They focused on monogenetic volcanoes that erupted only once and then became extinct, but this does not imply a "simple eruption.” Monogenetic volcanoes can have eruptive episodes with durations of hours to decades (or even centuries). The magma may evolve during eruptive episodes (fractionation, assimilation, arrival of different magma batches, tapping different mantle sources), changing styles of eruptions (effusive and pyroclastic), and quiescent cycles can complicate the eruption. In addition, a younger monogenetic volcano can overlap or be co-located with an older one.

Figure I-2. Graph of Total Alkali Silica (TAS) diagram of the basaltic rock names (Le Maitre, 1989). Compositions of basaltic rocks in the volcanic fields visited on the Desert Symposium field trip.
Effusive eruptions. Lava flows are the most abundant material formed during effusive eruptions. Effusive eruptions can begin as long fissure vents above the feeder dikes that soon focus to point sources. Both fissure and point vents develop lateral ramparts of successive spillover lava flows from the vent, which might include spatter and agglutinate. Effusive eruptions, especially Hawaiian type with sustained magma supply, can have lava fountains where the lava column separates into bursts and threads of lava that can form spatter and agglutinate cones when they fall back to the ground, although much of it can reform into flowing lava. From fountaining lava, fine particles of glass (congealed and quenched lava) can form ash, long threads of Pele’s hair, small lava droplets of Pele’s tears, as well as reticulite (a basaltic foam with 98–99 percent maximum porosity) that can be transported in the atmosphere to form fallout deposits. These forms of ash can locally form appreciable thickness but typically are not represented well in the stratigraphic record.

Effusive eruptions typically form small low-angle sloping shields near the vent, and if the vents are spatially centralized, they can form a shield that encompasses the entire volcanic field. Lava flows form a morphological spectrum from pāhoehoe to 'a'ā and blocky. Pāhoehoe flows have smooth surfaces and form long flowing flows or fans that progress by lava emerging from beneath the cooled crust through lobes and toes of the flow. 'A'ā flows typically have dense lava cores with very rough and rubbly surfaces of fragmented rocks (commonly referred to as clinker) that form at the flow surface as it is ripped apart by shear forces developed in the cooling and congealing lava. The upper surface of the lava is typically carried by the lava core, so that when the base of the lava stops flowing, the upper surface advances to the front of the flow where it forms a clinker base. Not all the fragmented and clinker base is from the overridden top of the flow because the cooling and congealing base also sets up the shearing that cause the basal fragmentation. Both pāhoehoe and 'a'ā can form levees that confine lava to channels that enable lava to travel long distances, and if the lava flows out from beneath the cooled roof, a lava tube is formed. Lava tubes can be reoccupied by later pulses of lava, or even lava from later flows. Both pāhoehoe and 'a'ā can form folds or ridges on the top of the flow that are typically arcuate and transverse to the flow direction of the lava. However, in pāhoehoe these folds are usually a few centimeters across and form a ropy structure, and in 'a'ā flows the folds can have 1–10 m wavelengths and can be several meters high.

Explosive eruptions. Explosive eruptions form a spectrum from magmatic, where all material is from the erupting magma (sometimes referred to as juvenile), to phreatomagmatic, where there is interaction of magma and water (groundwater, or surfiical water in lakes, rivers, or the ocean), to phreatic where heated water in a rock mass is triggered to explode and no magma is involved. Valentine et al. (2021) categorize phreatomagmatic explosive activity as producing maar-diatremes, tephra rings, and tephra cones. Maar-diatermes (White and Ross, 2011) have craters that cut into the pre-eruptive landscape and are surrounded by low-profile tephra rings. Tephra rings and cones include deposits of fallout, pyroclastic currents (pyroclastic surges, dominantly), and ballistic clasts, and normally contain abundant lithic clasts that can be dominant relative to juvenile clasts. Tuff rings and tuff cones have crater floors at or above the pre-eruptive ground level, and low- to high-angle slopes, respectively; they are commonly associated with shallow groundwater- or surface water-magma explosions. Juvenile material in such features commonly is highly altered to palagonite, a yellowish, hydrated basaltic glass.

Most eruptive centers in the Mojave region have had magmatic eruptions that formed scoria or “cinder” cones, and a few locations had phreatomagmatic eruptions that formed tuff cones (Miller and Buesch, 2022). The Cima field has tuff rings (some previously identified as maars), and Miller and Buesch (2022) suggest possible tuff rings at Ash Hill and Amboy. According to Valentine et al. (2021), phreatomagmatic landforms typically form ~10 percent or less of the Quaternary volcano types, so the few phreatomagmatic centers in the Mojave are consistent with this regional trend.

Topics for discussion on the field trip
During the trip a broad range of topics of studies conducted at the volcanoes visited will be discussed. The following list provides some highlights, but the details (including references) will be discussed at each stop.

- Pisgah, Dish Hill, Amboy, and Cima volcanic fields have been the sites of petrologic studies focused on what and where melting occurred in the mantle to form the magma, and what happened to the magma as it traveled from the depth of melting to the surface (for example, a chamber at some intermediate depth).
- Dish Hill and Cima volcanic fields have world recognized assemblages of mantle and crustal xenoliths (foreign fragments to the magma). The mineral composition textures, and rock structure permit modeling of temperature, pressure, and stresses at the depths of the xenolith sources, and how and where they were included into the magma.
Pisgah, Dish Hill, Amboy, Cima and Bicycle Lake volcanic fields have been the sites of studies of lava flows for paleomagnetic properties of lava flows including secular variations and magnetostratigraphy. The studies have been used to determine the ages of the lava, estimates of time between eruptions at a volcano, and how long-lived volcanic centers and fields have been active.

- Pisgah, Dish Hill, Amboy, and Cima volcanic fields were sites of studies of spectral analysis; originally 5-band (wavelength) Landsat, but now several forms of hyperspectral visual to thermal infrared imaging have been used for rock identification including alteration. These studies have been done to understand terrestrial and planetary volcanism.

- Some of the earliest terrain analyses of lava fields were conducted at Pisgah, and the Pisgah, Amboy, and Cima have a rich history of radar studies for landform and deposit morphology of vents, lava flow, sedimentary rocks, and aeolian processes.

**How to read the road log**

The road log contains the descriptions for the stops, and it will provide a detailed set of instructions to travel from stop to stop and see some drive-by sites along the way. There is road log for each day (D1 and D2). The format of the road log is that each stop (S), driving instruction (RL), or drive-by sighting (RL D) is assigned an identifier and there is a short description; for example, “D1 RL02 D Gate and small parking area”. The mileage driven from the last S, RL, or RL D is listed, and the UTM coordinates are listed: for example, “1.0 mi. UTM 11S 556553 3845813”.

Most of the roads traveled for the trip are paved, but at the Cima and Bicycle Lake fields the roads are dirt or in washes, and locally are sandy. The dirt roads at Cima, the morning of Day 2, have been in good shape, and the short drive along a wash does not need high clearance vehicles, but parts are sandy. In the afternoon of Day 2, the dirt road to the stops for the Powerline sequence and Bicycle Lake basalt field are along roads are infrequently maintained, so they vary in quality. There are stretches with loose sand and/or deep tracks and corrugated roadbed, and high-clearance vehicles are preferred. Additional road conditions to be aware of will be described when needed.

Drive-by locations have instructions for where to look, and it uses the “o’clock” convention. When driving or standing along a road at a stop, 12:00 o’clock (oc) is along the road in the direction of travel. If standing in the field, the speaker will define where 12:00 will be. To one’s right is 3:00, behind is 6:00, and to the left is 9:00. Where a feature such as a volcanic field is described the o’clock sweep can be either to the right or left and which way is defined: for example, 2:00 to 4:00 would be a sweep to the right.

On Day 1, the drive will be along the National Trails Highway (NTH). This road is the historic U.S. Route 66 that was established in 1936, and was one of the arterial highways from Santa Monica, California, to Chicago, Illinois. Because our trip will be traveling through the geologic past, it is appropriate to travel Route 66.
Please note that some stops are on private property and others are on Federal lands managed by Bureau of Land Management and National Park Service. Respect the land and leave it as you found it: no rock collecting, digging, or harming of plants and wildlife is allowed.

The desert can be a hostile place. Carry water and snacks, plan your clothing for the conditions, and always protect skin and watch for harmful plants and animals.

Day 1

STOP D1 S1-1. Overview of Day 1 trip and the Pisgah volcanic field

0 mi. UTM 11S 557308 3844930

Camp site, and stop D1-S1.

Overview

Day 1 of this two-day field trip will be along the National Trails Highway (a.k.a. Route 66) that follows the Barstow-Bristol trough, a long-lived NW-trending structural and topographic depression. This trip will focus on three basaltic volcanic fields: Pisgah (~23 ka), Dish Hill (~2.1 Ma), and Amboy (79 ka) (Wilshire and Trask, 1971; Wilshire et al., 1980; Phillips, 2003). The first stop is at the Pisgah volcanic field. South-southwest of the Pisgah field are the Sunshine and Lavic volcanic fields (Fig. 1-1). Based on petrology and erosional degradation, Sunshine is probably similar in age to the Pisgah field. The Lavic volcanic field is ~750 ka (Oskin et al., 2008), and similar to the Pipkin volcanic field in the Rodman Mountains farther west (~770 ka; Oskin et al., 2007). Previously, the Pisgah, Sunshine, and Lavic volcanic fields were grouped in the Lavic Lake volcanic field, but they are now considered to be individual fields.

Amboy and Pisgah were erupted in slightly different settings, where Amboy formed in the middle of a large basin and was deposited on very low slopes, and Pisgah was deposited on a topographic ridge that directed some of the lava flow paths. A great advantage of working with young deposits such as Pisgah and Amboy is that they provide insights into the volcanic processes that form them, and this knowledge can be used for interpreting the older deposits. We will visit Dish Hill to look at an older and more eroded field that was deposited across paleotopography developed on Jurassic and Cretaceous granitoids and has one of the world’s best representations of mantle xenoliths. The study of these xenoliths indicates the composition and structural history of the upper mantle and lower lithosphere. Because Amboy, Pisgah, and Dish Hill formed in such different environments, and the feeder dikes encountered such different rock types, we can begin to discuss these types of influences on the styles of eruption.

The Pisgah volcanic field location is in the eastern California shear zone (ECSZ), and there are two NW-striking faults of this shear zone nearby. The Lavic Lake fault is between the Sunshine volcanic field and Lavic Lake and was the fault on which the 1999, M\textsubscript{w} 7.1 Hector Mine earthquake occurred. Hector Mine is an open pit mine WNW of the Pisgah cone. Sylvester et al. (2002) mapped fractures and displaced rocks in the main Pisgah cone that were formed by shaking associated with the Hector Mine earthquake. On the west side of the valley, near the crest of the Sunshine and Lavic volcanic fields, is the Pisgah fault that is traced along the west side of a long NW-trending ridge where it crosses the National Trails Highway and I-40. Two of the lava flows from the Pisgah field flowed along the west side of the NE-trending ridge, and one flow was cut and disrupted by the 1992 M\textsubscript{w} 7.3 Landers earthquake.
The volcanology that we will examine today focuses on both effusive eruptions and explosive eruptions. Effusive eruptions are magmatic and erupt because of the flow of the magma to the surface. One type of effusive vent is a fissure that forms lateral ramparts with deposits of spatter and agglomerate. Fissure vents can become restricted to smaller areas (more of a point source). A second type is a dome or pyramidal structure formed by injection of lava that uplifted and deformed older flows, and fractures develop across the dome or pyramid where lava oozes out to form flows. A third type is lava lakes that can be hundreds of meters across and elevated (10 m or more) with ponded lava. Pāhoehoe or ‘ā‘ā can be emitted from fissures or the fractures at the edges of domes or pyramids, and at least at Amboy where these features are abundant, pāhoehoe typically forms lava lakes. Explosive eruptions can be magmatic where driven solely by the vesiculating magma to create Hawaiian or Strombolian eruptions that typically form scoria cones. Explosive eruptions can also be driven by interaction of water with magma to form phreatomagmatic eruptions that form tuff cones, tuff rings, or maars. At Pisgah, there are magmatic effusive and explosive features and processes. At Amboy, most eruptions were magmatic effusive and explosive, but there were also some possible phreatomagmatic explosive eruptions.

**Pisgah volcanic field**

Wise (1966) mapped the Pisgah volcanic field for NASA as an early characterization of a young basaltic field. This field has three overlapping scoria cones formed during explosive eruptions, and effusive eruptions formed fissure, dome, and pyramidal vents with extensive pāhoehoe and ‘ā‘ā lava flows (Fig. 1-2). The lava is basanite (Wise, 1969) and based on phenocryst content and minerals, there were three compositions of eruption phases ("p1", "p2", "p3" symbols are informal for this road log): (1) "p1" has microporphyritic olivine (smaller than 2 mm) and phenocrysts are scattered and rare, (2) "p2" has porphyritic olivine (2–3 mm) and plagioclase (2–5 mm) phenocrysts, and (3) "p3" has plagioclase phenocrysts larger than 10 mm and clots of olivine crystals about 5-6 mm across. Each eruptive phase consisted of a period of early explosive scoria cone-forming events followed by extensive effusive events. In each eruptive phase, the scoria cone is located NNW of the effusive vents, and in each successive eruptive phase the vent pairs moved slightly ESE of the previous vent pairs. Pāhoehoe flows from eruptive phases "p1" and "p2" flowed ~18 km to the WNW along a paleochannel.

The ~83 km² volcanic field was developed in the topographically low Barstow-Bristol trough, and the vents appear to have formed on a local highland and flowed to the SE, SW, and NW. A set of closely spaced samples from eruptive phase 1 yielded a ³⁶Cl age of 22.5 ± 1.3 ka (Phillips, 2003). The Pisgah field has been studied for igneous petrology (Wise, 1969; Glazner et al., 1991) as well as crustal contamination with Sr isotopes at phenocryst scales (Ramos and Reid, 2005), fault-induced damage to the scoria cone from the Hector mine earthquake (Sylvester et al., 2002), field testing the Rocky 7, the prototype for the Mars rover (Arvidson et al., 1998), and communications in lava caves by NASA (Belov et al., 2017). McCue and Green (1965) tested early digital...
Physiography of effusive vents and lava flows

The Pisgah field has effusive vents associated with (1) fissure eruptions and associated lateral ramparts, (2) dome or pyramidal shaped structures formed by injection of lava that uplifted and deformed older flows, and (3) possibly a few elevated ponded lava or lava lakes. Vent-lava features such as those identified at Amboy have not been identified in the Pisgah field. However, there are a few large platforms of flat-topped ponds of lava with 2–5 m relief of steep-sided marginal levees and margin-parallel tension fracture systems, which have collapse depressions and small domes, and lateral breakouts that formed secondary and tertiary small flows. Fissure vents can be hard to locate, but one formed during eruptive phase 2 that is 350 m long, 130 m wide, and 13 m high with an axial trough filled with blocks of 'a'a and pāhoehoe. Parts of this fissure vent were reoccupied during eruptive phase 3 with pāhoehoe to locally formed lava ponds and spill-over flows.

Because these vent structures are primarily lava flows, they can be very resistant to erosion, and recognizable from subtle topography. Most of the Pisgah lava field is pāhoehoe with lesser amounts of 'a'a, and only locally some 'a'a transitions into block lava; pāhoehoe flows typically have a longer runouts than 'a'a. The distribution of the flows results from flow across the ground surface, advancing the flow front, and can be enhanced by formation of lava tubes that deliver lava to the advancing flow front. In the topographic high areas of the vents, especially in pāhoehoe, lava tubes formed at all scales and were drained as the lava flowed down slope. In low-relief and distal areas, lava tubes can form but remain filled as the lava flows toward the advancing front.

Geochemistry

Based on samples from Pisgah, Sunshine, Lavic, Dish Hill, and Amboy volcanic fields, with four other volcanic fields that we will not visit (Fig. I-1), Wise (1969) demonstrated that the compositional trends in the alkalic basanite lavas were the reverse of that expected from differentiation of a single alkalic magma, and he hypothesized that the trends were caused by a succession of magmas generated from partial melting in the mantle. Wise (1969) concluded that changes in the basanite compositions with time are best explained as a rising melting zone, analogous to a mantle diapir. Glazner et al. (1991) used samples from Pisagh and Amboy, together with Dish Hill and Deadman Lake (Fig. I-1), to conclude that the geochemical trends resulted from interaction between mantle-derived magmas and preexisting mafic continental crust, and this interaction repeated in ponded chambers at density discontinuities in the crust. Glazner et al. (1991) also suggested that some of the geochemical trends resulted from cryptic contamination of Ocean Island Basalt (OIB)-like basalts.

Remote sensing

Pisgah volcanic field was one of the earliest fields identified by NASA for evaluating satellite and airborne remote sensing for rock identification and geologic mapping. Real Aperture radar (RAR) band K with HH and HV polarization, Synthetic Aperture radar (SAR)
band X and L with XHH, LHH, LHV polarization, and Seasat-A radar band L HH polarization data were available for the Pisgah volcanic field (Elachi et al., 1980), and they concluded the following:

1. The dominantly pāhoehoe units (first and third eruptive units) and the 'a‘ā lava of the second phase of eruption are distinguishable on LHV (L-band, horizontal transmit, vertical receive) and KHV radar images. The 'a‘ā–pāhoehoe contact can barely be discerned in the LHH image and cannot be seen in the XHH and KHV images. The first and third flow units cannot be separated on any of the images. In LHV, the second unit is brighter than its surroundings. The 'a‘ā of the second phase appears to be blocky at K-band (0.86-cm wavelength) and rough at L-band (25-cm wavelength).

2. The second phase can also be separated from the first and third phases in the Landsat image. The second phase is darker at visible and near-infrared wavelengths partly because of shadowing in the rough 'a‘ā lavas.

3. Discrimination between 'a‘ā and pāhoehoe lavas is also possible in the Seasat-A (LHH) image. The 'a‘ā areas to the east and southeast of Pisgah Crater are slightly brighter than the surrounding pāhoehoe, as in the LHH image acquired by aircraft. An area of hummocky pāhoehoe west of Pisgah Crater is also bright in the Seasat-A image. This is probably caused by the low slope angles of the hummocks furnishing normal specular reflection at the small incidence angle of Seasat-A (20°) and not at the larger incidence angles characteristic of the aircraft images.

4. The sand-covered western tongue of pāhoehoe is difficult to separate from its surroundings in K-band images, but X- and L-band clearly show the flow boundaries. This implies either that only the smallest scale (<1 cm) surface roughness has been masked by the wind-blown sediment or that longer wavelengths are able to penetrate the sediments. Aeolian material visibly masks portions of the flow in the Landsat image.

5. The cinder cone of Pisgah Crater is clearly delineated on the K- and X-band images because of their small depression angles and the resulting shadows, and on the Seasat-A image because, at the small incidence angle, surface slope change leads to a large tonal change in the image. It is barely discernible on the aircraft L-band image.

This summary of remote sensing focused on findings from Elachi et al. (1980), and subsequent papers refined and expanded upon these basic findings. It is this type of application of technique that launched the applications of Lidar and into terrain and object recognition used in autonomous vehicle navigation.

**The hike: Examining lava flow features**

This location on the SE side of the large scoria cone provides numerous options for examining the main eruptive vent areas of the Pisgah volcanic field. Today’s hike instead will focus on the effusive vents and lava flow features to the SE.

Before the hike begins, be aware that we will be walking on pāhoehoe and 'a‘ā that can be sharp, so wear study boots, long pants, and gloves. In many places, the terrain is rough and covered with unstable rocks. Slow is the way to go. Be sure about your footing. If you are not sure about a traverse, then don’t go. There are lava tubes in the pāhoehoe, but please do not enter them.

As a reminder, based on phenocryst minerals, size, and abundance, Wise (1966) identified three eruptive phases and the symbols (“p1”, “p2”, “p3”) are informally used in this road log.

1. “p1” is microporphyritic olivine (smaller than 2 mm) and phenocrysts are scattered and rare.
2. “p2” is porphyritic olivine (2-3 mm) and plagioclase (2-5 mm) phenocrysts.
3. “p3” has plagioclase phenocrysts larger than 10 mm and clots of olivine crystals about 5-6 mm across.

The traverse begins on the edge of the tailings terrace where it is the shortest walk to the highest exposed pāhoehoe. We will see rocks from the three eruptive phases identified by Wise (1966). We will go past “p3” pāhoehoe with various size lava tubes (some is almost shelly pāhoehoe), walk across “p2” ‘a‘ā and pāhoehoe that formed in an effusive fissure vent (called “the tongue” by Wise, 1966), look down on a “p3” pāhoehoe field with spatter cones, walk to the northern edge of a “p1”
pyramidal vent (called “the dome” by Wise, 1966), pass the “p2” vent for an ‘a’a flow, and return across “p3” pāhoehoe. Return to vehicles.

D1 RL01 D Return to Route 66 (National Trails Highway)
0 mi. UTM 11S 557308 3844930

D1 RL02 D Gate and small parking area – Alternative hike
1.0 mi. UTM 11S 556553 3845813

This location at a parking area adjacent to the gate to private property was established as an alternative hike to see the varieties of basaltic lava features. East of the road are excellent exposures of “p1” lava flows including ‘a’a flows with 20–40 cm high and 1–3 m wavelength ridges, tumuli, a mega-tumuli or fissure vent, pyramidal vents, a platform with collapse depression and small mounds, breached margins of platforms with effusive domes feeding pāhoehoe and ‘a’a flows, some of which are second- and third-derivative flows.

On the drive out, the road is along the western edge of the “p1” lava flows.

D1 RL03 Route 66 (NTH) Road – Turn LEFT
1.5 mi. UTM 11S 557303 3847937

D1 RL04 Pisgah “p1” flow and Pisgah fault
6.0 mi. UTM 11S 547973 3849757

The Pisgah lava at this location is “p1” (Wise, 1966) and it is cut and deformed by the Pisgah fault. This location was part of the Desert Symposium 2013 trip. Following is the entry from that trip.

“The Pisgah fault cuts the late Pleistocene Pisgah basalt flow (22.5 ± 1.3 ka from Phillips, 2003). The basalt flowed north onto the southern (Troy Lake) arm of Lake Manix within the period of the late Pleistocene before the lake drained. There are no pillow structures developed in the basalt, as one would expect if warm basalt flowed into lake water.”

D1 RL05 D Pisgah “p2”(w) contact with “p1”(e)
1.0 mi. UTM 11S 546241 3850209

STOP D1 S1-2. Distal lava flows of the Pisgah volcanic field

D1 S1-2 Pisgah flow “p2” (short hike)
0.4 mi. UTM 11S 545600 3850330

D1 RL06 Depart Stop 1-2 and retrace your route to the east

D1 RL07 Route 66 (NTH) – Pisgah “p1”(w)-“p2”(e) contact
7.6 mi. UTM 11S 557479 3847828

The road crosses the “p1” and “p2” contact at about a 60° angle. The “p1” is to the SW in the lower ground, and the “p2” is to the NE in the higher ground.

D1 RL08 D View W of Pisgah, Sunshine, Lavic volcanic fields – Road turns LEFT
4.6 mi. UTM 11S 563663 3843898

Just before the road turns left, the view to the west is of the Pisgah, Sunshine, Lavic volcanic fields and Miocene basalts.

O’clock Topic
5:30 to 3:00 Pisgah volcanic field
4:00 to 3:30 Lavic volcanic field
3:30 to 2:45 Sunshine volcanic field. (Sunshine scoria cone is ~3:20)
2:45 to 2:00 The dark rocks low in the hills are Miocene basalts

D1 RL09 D Bridge across freeway – Road bends RIGHT; continue on Route 66
0.3 mi. UTM 11S 563790 3844252

D1 RL10 View of Cady & Bristol Mountains and Crucero Rd, Ludlow – Turn RIGHT and pass under the freeway
8.1 mi. UTM 11S 576601 3843059

Best views are while on the road just before Ludlow. View of Cady & Bristol Mts; Broadwell Mesa, Broadwell Lake.

O’clock Topic
7:00 to 9:00 Cady Mountains
9:00 to 12:00 Bristol Mountains
9:00 Broadwell Lake
10:00 to 10:30 Broadwell Mesa basalt field

D1 RL11 Route 66 (NTH) – Turn LEFT, leaving Ludlow
0.2 mi. UTM 11S 576621 3842729

D1 RL12 D Ash Hill volcanic field – On left, possible low-relief tuff(?) cone
4.4 mi. UTM 11S 583316 3841280

D1 RL13 D Ash Hill volcanic field – On right, possible cone remnant on ridge crest
2.7 mi. UTM 11S 587221 3839330
Three volcanoes form the Dish Hill cluster north of Route 66. Dish Hill is a 175 m high scoria cone, Hill 1933 is a 120 m high scoria cone, and Hill 1069 is scoria cone partially buried by alluvium and has 43 m relief. Both Dish Hill and Hill 1933 were deposited on Jurassic and Cretaceous granitic rocks, and the relief described is above the bedrock. All three volcanoes erupted basanite lava, and many lava flows and pyroclasts contain mantle xenoliths and kaersutite megacrysts (Wise, 1966; Wilshire and Nielson-Pike, 1986). Basanite lava in the Mojave Desert occurs at Dish Hill, Deadman Lake, Lead Mountain, Pisgah, Sunshine, Lavic, and Pipkin volcanic fields (Wise, 1969). Dish Hill is early Pleistocene in age based on three dates: a 2.1 ± 0.2 Ma fission-track age determined on apatite in a granitic inclusion (C.W. Naeser in Wilshire and Trask, 1971), a 2.03 ± 0.12 Ma K-Ar age on a selvage (fracture-fill) amphibole (M.A. Lanphere in Wilshire et al., 1980), and a 1.9 Ma K-Ar date on amphibole (Wilshire and Nielson-Pike, 1986). The Dish Hill cluster might be part of a larger contemporaneous volcanic field that includes Deadman Lake volcanic field (Howard, 2022).

Wise (1966) and Wilshire and Nielson-Pike (1986) mapped the three volcanoes (Fig. 1-4). Dish Hill is a scoria cone with five lava flows, the longest of which (~1.6 km) emanated from a breach in the west side of the cone and rafted parts of the cone, and locally dikes cut the scoria cone and might have fed small flank effusive vents (Fig. 1-5). Hill 1933 is a scoria cone with four lava flows, the longest of which (~2.5 km) emanated from a breach in the west side of the cone and carried with it rafted parts of the cone. The scoria cone has upper pyroclastic deposits referred to as agglutinate where clasts are sintered into a hardened rock, and the lower slopes of the cone are volcanic (pyroclastic fallout) breccia with lapilli to block clasts with various amounts of tuffaceous matrix (some altered to palagonite). The lava flows at Dish Hill (and Hill 1933) were relatively fluid, up to 7.6 m (25 ft) thick, have 0.7-1.7 m (2–5 ft) thick vesicular zones at the top and bottom of the flow, a sparsely vesicular core, and all flows contain xenoliths (Wise, 1966). The tops and bottoms of the flows have up to 20 percent glass resulting from rapid chilling. The crystallized lava and pyroclasts are extremely fine grained (<0.1 mm) with microphenocrysts (<0.5 mm) of olivine, and groundmass grains are olivine, titanomagnetite, ilmenite, and plagioclase. In the cores of lava flows, there are small patches of phillipsite-chabazite intergrowth that Wise (1966) interpreted as initial crystallization of the glass during cooling of the lava that formed zeolites rather than feldspathoids (that is, they are not deposited from groundwater).
Dish Hill lavas have ultramafic xenoliths and phenocrysts derived from xenoliths that represent rocks from various depths (pressures and temperatures) in the mantle, and textural studies provide insights to their structural deformation.

- **Xenoliths**—Each type of a xenolith implies a composition (especially the ultramafic rocks, which different ratios of olivine, orthopyroxene, and clinopyroxene), and these rocks can have a specific mineral (such as Cr-diopside) as a defining adjective. Xenoliths include: peridotite and minor amounts of partially fused granite (Wise, 1966), and peridotite, herzolite, wehlrite, harzburgite, pyroxenite and partially fused granite (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009).

- **Minerals**—Xenocrysts include: titanaugite and kaersutite (Wise, 1966), and kaersutite (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009). Fracture fillings include: kaersutite, phlogopite (mica), and rare apatite and opaque oxides, and very rare plagioclase (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009).

- **Structural features**—Structural features in the xenoliths include granoblastic texture, kink banding of olivine, foliation from mylonitization and recrystallization (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009), sharp contacts of pyroxene-rich and olivine-rich rocks, polished fractures (by fusion of amphibole rather than abrasion polishing), up to six faceted surfaces on a xenolith (the breaking of much larger blocks), and veins of different compositions crossing or offsetting foliation or kink bands in olivine (Wilshire and Trask, 1971). Most textures are microscopic, but macroscopic features include the polished fractures and faceted surfaces.

- **Geochemical analyses**—Geochemical analyses published (or referred to) in papers include major oxides, minor-trace elements, and Rb, Sr, H, O isotopes (Wilshire and Trask, 1971; Wilshire et al., 1980, 1988) invoked a rising diapir model, and Luffi et al. (2009) invoke sampling of tectonically subcreted and imbricated oceanic lithosphere emplaced during low-angle subduction.

The rocks at Dish Hill and the processes that formed them have been analyzed in numerous papers on petrologic development of mantle melts, characterization

Figure 1-5. Photograph of the west side of Dish Hill. The main cone breech exposes (1) internal stratigraphy of the cone, (2) vent-filling lava pond, (3) rim-spllover lava flow, and (4) the 1.6 km long lava flow capped by rafted scoria cone pieces (behind the tops of the telephone poles). Photograph by David Miller (2021).
of mantle and lower lithospheric xenoliths and inclusion in the magmas, and the tectonic evolution of the Mojave lithosphere (Wise, 1966; Wilshire and Trask, 1971; Wilshire et al., 1980; Wilshire and Nielson-Pike, 1986; Wilshire et al., 1988; Luffi et al., 2009). For more descriptions of the volcanic rocks and xenoliths in the Dish Hill cluster, see the summary paper by Bridenbecker (2022).

At this stop, there is no guided walk. Individuals can explore on their own to look for xenoliths or at the tephra deposits, but don't wander too far.

D1 RL14 Return to Route 66 (NTH) – Turn LEFT
0.6 mi. UTM 11S 596755 3829173

D1 RL20 View of Amboy main vent-lavas
5.2 mi. UTM 11S 604892 3826912
View of Amboy main vent-lavas, Amboy cone
O’clock Topic
0:45 Amboy scoria cone
1:10 to 2:10 Amboy vent lavas at the apex of the broad low-relief basaltic shield volcano

D1 RL21 Amboy basalt pressure ridges at Route 66 (NTH)
3.6 mi. UTM 11S 610517 3825558

D1 RL22 Route 66 and Amboy Crater Road & parking – Turn RIGHT
1.2 mi. UTM 11S 612411 3824962

STOP D1 S3-1. Parking lot in Amboy volcanic field, center of the field
D1 S3-1 Amboy visitor overlook—short hike
0.5 mi. UTM 11S 611798 3824614

The Amboy volcanic field is a 70 km², low-relief (0.3-1.8°) shield dominated by pāhoehoe flows as long as 6.5 km that originated at effusive centers clustered near the apex of the shield, and there is a well-formed scoria cone on the northeast slope with associated small effusive vents near the base (Parker, 1963; Greeley and Iverson, 1978) (Fig. 1-6). This field has extensive pāhoehoe with complex morphology of intermingled flows and lobes, and formation of tumuli and pressure ridges. Greeley and Iverson (1978) identified flat topped features termed “platforms” and “vent lava”. Platforms occur throughout the field, can cover large areas, and are 1-5 m in relief. Vent lavas are also flat topped, but the tops can form broad depressions, cover areas hundreds of meters across, and are ~10 m in relief. Both platforms and vent lavas can have collapse depressions, and Greeley and Iverson (1978) used the broad depression of the top and the size of collapse depressions to infer that a significant amount of lava was withdrawn from the vent area and flowed back down the conduit. Parker (1963) mapped the largest vent lava and referred to it as “the plateau”, and recognized the collapse depressions, and also numerous chaotic jumbles of basalt blocks that he interpreted as possibly phreatic deposits. Recent mapping indicates a few possible, small low-relief tuff rings or maar deposits.

Amboy volcanic field has been the site of volcanic field morphology (cinder cone, lava flows) and aeolian process studies, including by NASA for earth and planetary applications (Greeley and Iverson, 1978). The ~70 km² volcanic field was developed in the Bristol Lake basin within the ECSZ. A lava flow in the field has a 79±5 ka ⁴⁰K⁴⁰Ca age (Phillips, 2003), and the basalt flows were deposited in the distal alluvial fan and playa facies of a ~4 Ma depocenter (Rosen, 1991; see later discussion). Types of studies include igneous petrology (Parker, 1959; Wise, 1969; Glazner et al., 1991), geologic mapping of basaltic rocks and determining rock properties for analogs of planetary geology (Greeley and Bunch, 1976), formation of craters in basalt flows for planetary studies (Greeley and Gault, 1979), remote sensing with aerial photography...
and some of the early Landsat and airborne radar (Elachi et al., 1980; Arvidson et al., 1998), and aeolian processes (Greeley and Iverson, 1978). Amboy was also used for field testing Rocky 7, the prototype for the Mars rover (Muckenfuss, 2001). There were even rock mechanical properties and penetrator drop tests where a penetrator was dropped as a free-fall projectile from an altitude of 2500 m to impact the basalt flow at a velocity of 213 m/sec (Blanchard and others, 1977). Geologic maps of the field by Parker (1963), Hatheway (1971), and Greeley and Iverson (1978) focused on morphology of the cinder cone and lava flows, and are summarized here. Some of the oblique aerial photographs from Greeley and Iverson (1978) depicting the lava flow features are included as well. From a cultural and anthropomorphic perspective, there was a Mohave Nation story about the formation of Amboy Crater by AH MOTT KAH PEE THOYAH, who hunted with fire (Bridenbecker, 2022). Amboy appears to have had a couple of prankster events when people lit tires to simulate an eruption; however, these events are not well documented and might be more urban legend than fact (Muckenfuss, 2001; McShane (2011).

Scoria cone

Parker (1963) described the ~75 m relief main cinder (scoria) cone and crater with internal conelets (Fig. 1-7). Lithologic features on the cinder (scoria) cone include variations in grain size of pyroclasts (up to bombs), shapes of bombs that include angular, and almond-shaped and ribbon (that indicate ejected fluid-like lava fragments), locally agglutinate (or welded spatter), and locally lava flows. From the distribution of these features, a seven-stage eruption history was reconstructed. It began with the eruptions to form the main cone and ended with lava flows emanating from the base of the cone.

Cinder cone event history
1. Early eruptions were explosive with many fluid bombs.
2. Deposition of agglutinated aggregate of basaltic bombs on the rim and western flank.
3. Formation of an inner conelet by mild explosive eruptions. [Glazner et al. (1991) interpreted these deposits as a directed lava fountaining event.]
4. Breached western cone walls by a sideways-directed explosion or by the lava flow that occupies the breach.
5. Formation of another inner conelet.
6. Eruptions from the cone crater ended.
7. Possible effusive eruptions of fluid lava from the base of the cone.

There are three alternate interpretations of these events. (1) Events 1 and 2 are consistent with typical growth of a cone and represent different deposits based on what type of material fell back and was deposited from the eruption column. (2) Glazner et al. (1991) interpreted event 3 deposits as a directed lava fountaining event. (3) Glazner et al. (1991) state the cone was deposited on lava flows, but did not see lava flows emerging from the base of the cone. Glazner et al. (1991) stated that the relative ages of widespread lava flows and the scoria cone are not known.

Effusive vents and lava

Effusive vents and lava flows have numerous shapes that form intricate patterns in the volcanic field (Fig. 1-6, 1-8). Effusive vents are in the topographic high areas near the apex of the shield, and they form a north-northeast trending cluster of vents. The variations in lava flow morphology occur throughout the volcanic field.

Parker (1963) identified numerous morphologic features in the lava flows throughout the volcanic field, resulting from detailed descriptions and measurements. Individual pāhoehoe flows typically are 0.3-4 m thick with many branching toes and lateral breakouts, some with blocky margins such as snouts of flows, and typically there are well-developed ropy structures, some channels with levees, and no lava tubes were identified. Tracing individual flows was difficult because of similar compositions of the flows and discontinuous exposures disrupted by aeolian sand deposits. Tumuli, or pressure

Figure 1-7. Aerial photograph of Amboy scoria cone with inset conelets and lava flow from the breach. Photograph by David Miller (1980).
ridges, are up to 30 m long and 15 m in relief, and typically have a fracture along the long axis. Some tumuli are only several meters wide and long and a few meters in relief. Locally, domes are up to 30 m in diameter, up to 3 m in relief, typically have ~3 m diameter depressions at the top, and have one or more radial fractures.

Large flat-topped areas with steep sides typically have circular to oval depressions up to 50 m across (commonly bounded by collapse fractures). Some of the steep-sided flat-topped (parental) flow units have secondary (derivative) flows that emerged from the base of the steep flow margin, and these secondary flows can themselves have derivative flows (Parker, 1963). One area about 550-850 m wide and 1,500 m long, referred to as "the plateau", includes two different types of features (Parker, 1963, in his Fig. 6). In the northern half of the plateau, there are 14 piles of chaotically arranged basalt blocks (referred to as jumbles) that are 3-13 m in diameter and 1-4 m in relief; there are no fusiform lapilli or blocks, and some blocks are oxidized to red (there is one jumble in the southern area). In the southern half of the plateau, there are 12 bowl-shaped depressions, mostly 8-100 m in diameter and 1-13 m deep, and some depressions have raised rims of basalt blocks. There are two large depressions, the largest of which is ~300 m across. Parker (1963) proposed two explanations for the plateau with jumbles and collapse pits: (1) a primarily effusive vent with minor explosive activity, and (2) lava from a vent located elsewhere flowed onto wet sediments on the floor of a playa resulting in steam being trapped beneath the advancing flow and formation of phreatomagmatic explosions (from a rootless vent).

The map by Parker (1963) displays lava flow directions and the long axes of pressure ridges, and although there are general flow directions away from the area of the main cone, many areas have locally radiating flow directions. A schematic cross-section of the field shows a central main dike beneath the cinder cone, minor splays of the central dike to small satellite vents for lava flows, and a separate dike that fed lava flows, with the inference that some lava flows might be derived from distributed dike-fed vents (Parker, 1963).

Greeley and Iverson (1978), building on the mapping by Parker (1963) and Hatheway (1971), refined descriptions and names of some lava flow morphologic and possible vent-related features (Fig. 1-8). As described by Greeley and Iverson (1978), the lava field consists of large areas of undifferentiated pahoehoe flow units that form a relief of 2-5 m with tumuli and pressure ridges, no lava tubes, with only a few lava channels. There are large lava flow features described as (1) the oldest flow areas (not separated) with numerous "collapse" depressions (circular pits up to 10 m diameter several meters deep), (2) platform units, (3) "vent" lavas, and (4) pyramidal structures. The "collapse" depressions could form by collapse of the crust above fluid lava (by inference, removal of support by the fluid lava as it flowed elsewhere), or by inflation of the still plastic crust (by inference, influx of more fluid lava beneath the crust). Platform units are isolated areas of uniform basalt that have relatively flat surfaces with steep sides, are up to 350 m by 1,600 m with a typical relief of ~2-5 m (but some up to 8 m) relative to the surrounding lava flows, have margin-parallel tension fracture systems interpreted as the boundary between the ponded lava and the cooler levee, and many have numerous collapse depressions (similar to some flat topped and "plateau" areas of Parker, 1963). The platforms appear to be stagnant parts of lava flows (ponds) that solidified in place with little lateral movement after emplacement.

"Vent" lavas have many characteristics of the platforms, but are inferred to be colocated with vents (Fig. 1-8). The vent lavas have relatively flat tops and steep sides, up to 900 m by 1600 m across with a relief of 10-30 m relative to the surrounding lava flows, and the lava lakes cooled in relatively stagnant ponds. In one vent lava, the lava lake within the rims forms a broad depression with a relief of ~12 m below the rim. Vent lavas typically have one or more collapse depressions or craters, and as with other collapse depressions, they might have formed from failure of the newly formed crust by removal of lava as it

Figure 1-8. Aerial photograph of Amboy basaltic lavas at "vent lavas" (center left), scoria cone (upper right), and platforms (distributed).
flowed elsewhere. The largest collapse crater (in the large vent-lava area) is 150 m diameter, is centered in a 300 m diameter depression, and has an obvious rocky ledge forming the rim that appears to be slightly elevated above the surrounding deposits. Vent lavas can have one or more small (~2 m high) mounds of rocks or lava domes. Vent lavas typically have margin-parallel tension fracture systems resulting from sagging of the central low-relief ponded lava (lava lake) and possible removal of lava by drain-back flow down the vent, or breakout flows from the base of the margin forming secondary (or derivative) flows. Pyramidal structures are pyramidal in profile and triangular, trapezoidal, or pentagonal in map view with three to five radial fractures (or fracture sets). Greeley and Iverson (1978) identified two pyramidal structures that have up to 18 m of relief and are about 225 m × 350 m at the base. As with the much smaller versions of tumuli, the pyramidal structures are interpreted as upwardly domed rocks resulting from injection of lava beneath a well-developed lava crust.

Geochemistry

Parker (1959) did initial geochemistry in the context of magmatic differentiation. Wise (1969), building on his work at the Pisgah field, included a lava flow sample from Amboy in his examination of basanite volcanic fields in the Mojave. He demonstrated that the compositional trends in the basanite from these fields were the reverse of that expected from differentiation of a single alkalic magma, and interpreted a succession of magma generated from partial melting in the mantle. Wise (1969) concluded that changes in the basanite compositions with time are best explained as a rising melting zone, analogous to a mantle diapir. Hughes (1986) sampled in a variety of locations in the Amboy field, and worked out the eruption sequence from the scoria cone. He confirmed and refined the geochemical trends, and concluded they resulted from a sequence of partial melting at successively shallower depths. Glazner et al. (1991) concluded that the geochemical trends resulted from interaction between mantle-derived magmas and preexisting mafic continental crust, and this interaction repeated in ponded chambers at density discontinuities in the crust.

Remote sensing for mapping and rock identification

Amboy volcanic field was one of the earliest fields identified by NASA for evaluating satellite and airborne remote sensing for rock identification and geologic mapping. Synthetic aperture radar (SAR) band X and L XHH, LHH, LHV and Seasat-A radar band L data were available for Amboy Crater and lava field (Elachi et al., 1980), and they concluded the following:

1. The X-band image has sufficient resolution for recognition of many geomorphic types within the flow. The vent lavas and some of the remnant platform units that have relatively smooth desert pavement surfaces are prominent. The undifferentiated flows are very irregular, as the many shadows and bright radar returns indicate. The scoria cone can be seen in the images because the large incidence angle of the X-band system accentuates topographic features.

2. Large subsidence fractures around the peripheries of the vent areas and collapse depressions are visible in the X-band image as dark lines, because aeolian material has filled the fractures and formed smooth surfaces. The piles of chaotic boulders on the main vent plateau are visible as bright point returns.

3. Although the L-band images have a lower resolution than the X-band image, the scoria cone, vent lavas and remnant platform units are visible whereas subsidence features and the piles of chaotic boulders are not. The oldest platform unit on the east side of the flow, however, is not obvious on any of the images and appears to be buried by deposits of Bristol Lake.

4. The aeolian material masking portions of Amboy lava field is clearly shown in the Landsat image by the lighter tone of the lava field as compared to Pisgah lava field. The sediment-free wind shadows are also easily identified as dark streaks trailing southeastward from topographic obstacles. These areas are visible on the radar images as slightly brighter returns because the surface roughness is less attenuated.

This summary focused on the findings of (Elachi et al., 1980). Subsequent papers refined and expanded upon these basic findings.

Return to Route 66 (NTH)

D1 RL22 Route 66 (NTH) – Turn RIGHT

0.5 mi. UTM 11S 611798 3824614

D1 RL23 Route 66 (NTH) and Amboy Road – Turn RIGHT

1.1 mi. UTM 11S 614108 3824441

D1 RL24 Amboy Rd – Curve RIGHT

0.6 mi. UTM 11S 615100 3823849

STOP D1 S3-2. Amboy volcanic field, center of the field

D1 S3-2 Amboy Rd – Drive past stop, U-turn, and return to Park

2.5 mi. UTM 11S 615270 3820143

This location is ~6.4 km ESE of the vents near the top of the broad Amboy shield, where the easternmost basaltic flows are overlain by playa deposits of Bristol Dry Lake. This provides an excellent setting to examine lava flow morphology and to evaluate the paleogeography and paleotopography at the time the Amboy volcanic field formed.
Bristol Dry Lake has nine boreholes, many <307 m deep; CAES #1 is 529 m deep and CAES #2 is 537 m deep (Rosen, 1991). A tephra layer in a basin-center core at 513 m depth has a tephrochronologic age of approximately 3.7 Ma (Rosen, 1989). From mapping, pits and trenches, and the core in these boreholes, Rosen (1991) reconstructed the stratigraphy and depositional history of the basin for the last 4 m.y. The sedimentary facies in the basin (from alluvial fan to basin-center) consists of (1) medial to distal fan deposits along the margins, (2) play margin sand float and wadi system, (3) saline mud-flat, and (4) salt pan at basin center. The present land surface in the alluvial fans has a low gradient (1.8 m/km; ~0.10°), and the fans tend to be slightly steeper on the north side of the basin compared to those on the south. There is no evidence for how far the alluvial fan and playa contact might have extended to the west, but because that would be along the topographic trough, any changes in slope are anticipated to be very gradual.

Borehole CAES #1 is ~1 km W of this location, and cuttings of basalt were from depths of 0-15 m (possibly 20 m). The exposed lava flow thickness is ~5 m, and if the base of the basalt was at 15 m, this indicates ~10 m of basalt is covered by the playa deposits. The next closest wells are 4.5-5.5 km to the south and east and don’t encounter basalt, so it is possible that the lava flows extend a little farther to the east than what we see at present.

Morphologically, lava flows in this area are pāhoehoe with ropy surfaces and tumuli and platforms (Greeley and Iverson, 1978). At the field trip stop, the platform and features of the platform are highlighted by the contact with the onlapping playa and surficial sediments (Fig. 1-9). The platform has a flat (or slightly arched or sagged) top, typically 1-5 m thick, has local collapse depressions, and steep margins with tension fractures along the rim. Where the dark lava is closest to the road, as many as three thin flows are exposed (possibly small, stacked lobes?), and in the topographically low areas around the high-standing lava flow, locally there appears to be small exposures of the tops of buried flows, implying a greater thickness of lava here than is exposed. Throughout most of this image, the darkest gray top of the flow has <1 m of relief, even way across the image to areas separated from the rest of the flow by wide embayments. This gives the impression that this was a ponded lava flow. Collapse depressions are attributed to withdrawal of lava from beneath a partially cooled crust across the pond surface; however, typically the depressions are not aligned as along a lava tube, and this led Greeley and Gault (1979) to suggest small breakouts along flow margins to form small secondary flows. Parts of the perimeter of the flow appear to be geometrically simple, but parts appear to be highly convoluted. Is this perimeter that of an advancing flow across a near horizontal surface such as a playa, or might parts represent collapse of the pond wall and headward propagation of the foundering pond surface?

Near the road, there is a good exposure of the top of the basalt lava flow that is overlain by silt-sand capped by a 1-2 clast thick pavement. As described previously, there is good evidence for burial of the lava flow by deposition of playa and surficial deposits that abut the margins of the flow. However, at this location, there is no evidence that the playa deposits were ever at the higher elevation of the top of this lava flow.

**Return to cars and proceed north**

D1 RL23 Amboy Rd and Route 66 (NTH) – Turn RIGHT, cross the railroad tracks with caution, and proceed through the historic town of Amboy; continue east 3.1 mi. UTM 11S 614108 3824441

D1 RL25 D Route 66 (NTH) and Kelbaker Road, Turn LEFT on Kelbaker Road 6.6 mi. UTM 11S 624554 3825115

To the east, Route 66 crosses the southern Marble Mountains. At the south end of the range is the site for the airborne Infrared spectral mapping described by Adams et al. (2022).
D1 RL26 D Marble mine on the west and Iron Hat Mine to the east
1.3 mi. UTM 11S 624305 3827125
O’clock Topic
2:00 Iron Hat mine in Iron Hat Canyon, Marble Mountains.
9:00 After years of inactivity, the high-purity calcite mine west of the road in the south Bristol Mountains has recently expanded operations.

D1 RL27 D Panorama of the Marble Mountains with deformed Paleozoic rocks and Miocene volcanic rocks on the skyline
2.7 mi. UTM 11S 6227711 3831214
O’clock Topic
3:00 A short segment of the Bristol-Granite mountain fault zone is exposed near the base of the lowermost dark exposures.

D1 RL28 D Northern Marble Mountains and Miocene Lost Marble paleovalley
2.7 mi. UTM 11S 6227711 3831214
O’clock Topic
2:00 Northern Marble Mountains where a high flat mesa (~7 km away) is capped by cliffs in the 18.8 Ma Peach Spring Tuff with the informally named Lost Marble gravel forming the slope beneath the tuff, and these rocks are in the west-trending Lost Marble paleovalley (Lease et al. 2009). The informally named Castle basalt forms the steep cliffs, and the informally named Brown Butte dacite forms the low ground (Glazner and Bartley, 1990).

Kelbaker Road crosses Winston Wash, which contains exposures of interbedded eolian and fluvial sediments deposited when the Kelso dunes aggraded across this part of the piedmont and blocked ephemeral drainages during the Pleistocene to Holocene transition (Sweeney et al., 2020). To access the outcrops, hike west along the wash for about 1.4 miles (2.3 km) to see dune-blocked channel fill that includes beds of sand, silt, and clay, as well as gravely alluvial fan sediments mantled by stabilized dunes. Today, the stabilized dunes allow ephemeral drainages such as Winston and Cottonwood washes to flow unimpeded in their channels through the dunes, where they meet up with Kelso Wash downstream.

D1 RL32 D Intersection with road on right near Kelso Visitor Center (closed) – Continue north
5.0 mi. UTM 11S 622790 3875202

D1 RL33 D Overview of Cima volcanic field
10.8 mi. UTM 11S 616956 3890057
O’clock Topic (Rotate left)
2:00 to 11:00 Cima volcanic field
12:00 Scoria cone V7, and the camp site is at the eastern base of the cone.

D1 RL34 Kelbaker Rd and Aiken Mine Road – Turn RIGHT
4.1 mi. UTM 11S 610167 3890932
The road was in good shape in November 2021. Driving on the dirt roads – slow down to 35 MPH (or less).
Near where the rocks on the north side of the road are at the wash, the road crosses the wash, and it might be sandy.

D1 RL35 Aiken Mine Rd and dirt road – Turn LEFT
3.1 mi. UTM 11S 613247 3897102
Campsite is in clearing ~250 ft from the road. This will be Stop 2-1 tomorrow.
Please follow the Leave No Trace approach to camping.
Only camp on previously disturbed ground. No collecting of wood for fires. We will need to confirm if a campfire will be allowed.

Day 2

DS22 Day 2 Stop 1 – Cima volcanic field
Welcome to the Mojave National Preserve, which was designated in 1994 and is managed by the National Park Service. Many geologic, botanical, and biological studies were conducted in the 1980s and 1990s to evaluate the natural resources of this area and subsequently to supporting the establishment of the Preserve. Much of the Preserve is designated as wilderness areas, including some of the areas that we will visit. Our trip has a Special Use Permit for the camping site on Sunday night and traveling
to and hiking around the three stops on Monday. Please do not disturb plants and soil (camp only on already disturbed ground), and please stay on the roads or the existing tracks in washes. Collecting is not allowed.

STOP D2 S1-1 Overview of the Cima volcanic field

0.0 mi.  UTM 11S 612163 3895035

Camp site and Stop D2 S1-1.
The Cima volcanic field is ~300 km² in size and has at least 71 vents of basaltic rocks resulting from different types of explosive and effusive eruptions from 7.5 Ma to 12 ka (Turrin et al., 1985; Wilshire et al. 2002a,b,c,d; Phillips, 2003). The Cima volcanic field is a national treasure for the geologic sciences, providing information on (1) volcanology, (2) geochemistry and formation of magmas, (3) compositions of the mantle from xenoliths brought up with the erupted magma, (4) paleomagnetism, (5) geochronology and correlation to the geologic timescale, (6) the evolution of weathering and erosion of scoria cones, (7) geomorphic processes such as the formation of desert pavement, and (8) applications of remote sensing (e.g., Arvidson et al., 1993; Dohrenwend et al., 1986; Elachi et al. 1980; Farmer et al., 1995a,b; Luffi et al., 2009; McFadden et al., 1984; Turrin et al., 1984; Wells et al., 1994; Wilshire et al., 1988, 1991, 2002a-d). This field trip will focus on the southwestern part of the Cima field, with three stops during the trip (S1 to S3) and short discussions about four alternative trips (S1-AS1 to S1-AS4) that are described in Gans (2022) (Fig. 2-1).

The lavas of the Cima field range widely in age. K/Ar ages range from 7.5 Ma to 60 ± 30 ka (Turrin et al., 1985; Wilshire et al., 2002a, b, c, d). Turrin and Champion (1991) dated one flow by 40Ar/39Ar at 119 ± 38 ka. Recent work on the youngest flows by cosmogenic methods with 3He yielded an age of 13 ± 3 ka (Wells et al., 1995), and with 36Cl yielded an age of 12 ± 2 ka (Phillips, 2003). Volcanism occurred during three periods of time and in different areas: (1) 7.5–7.0 Ma flows restricted to the southeastern part of the field, (2) 6.2–3.2 Ma flows spread over the entire field, and (3) <1.0 Ma flows restricted to the south half of the field.

Compositions of the volcanic rocks include mostly trachybasalt (47–52 SiO₂ wt% and 5.0–6.6 Na₂O+K₂O wt%), minor basalt (47–52 SiO₂ wt% and 4.9–5.0 Na₂O+K₂O wt%), rare basanite (52 SiO₂ wt% and 5.2 Na₂O+K₂O wt%), and minor basaltic trachyandesite (51–55 SiO₂ wt% and <6.6–7.6 Na₂O+K₂O wt%) (Katz and Boettcher, 1980; Wilshire et al. 1988; Farmer et al., 1995a). These analyses are from lava flow samples collected across the volcanic field.

The Cima volcanic field has scoria cones and tuff cones that resulted from pyroclastic eruptions and lava flows that resulted from effusive eruptions. The three forms very different vents and deposits. These types of eruptions range from relatively passive effusive eruptions where lava flowed from the vent, to moderately to highly
explosive eruptions that formed from magmatic and phreatomagmatic eruptions producing scoria cones and tuff rings, respectively.

Scoria cones are the most abundant vents in the Cima field and consist of juvenile coarse ash (0.5-2 mm), lapilli (2-64 mm) and bombs (>64 mm) that formed during Strombolian eruptions where the pyroclasts were lofted in the eruption column and fell back to form the cone around the vent. Scoria is moderately to highly vesiculated basalt (the field name) and is often referred to as cinder. Dohrenwend et al. (1986) described beds of scoria as well-defined with bedding contacts sloping parallel to the cone surface and consisting of coarse sand to 15 cm clasts that are not compacted but are tightly packed with bombs (typically fusiform, indicating viscous fluid) up to 2 m long. Grain fabric (long axis) is typically parallel to bedding and implies deposition as fallout. The beds can have normal, reverse, or compound size grading that implies changing conditions in the eruption column and potentially changing distributions of where the main deposition occurred around the vent. In the upper parts of the scoria cones, agglutinate or welded spatter mantles the upper part of the cone and is interstratified with nonconsolidated deposits. The scoria cones have mean cone heights between 50 and 155 m, mean widths between 400 and 915 m, summit craters up to 45 m deep and 330 m wide, and outer slope angles of 22° to 30° (Dohrenwend et al., 1986).

Dohrenwend et al. (1986) identified four tuff rings at the Cima volcanic field, including V12 and V13 (Fig. 2-1). Wilshire (2002d) described a maar deposit in the southeastern part of the field, and there are additional tuff rings in the southwestern part of the field that are described in this field trip and in Gans (2022). Phreatomagmatic eruptions result in tuff cones, tuff rings, and maars, and the differences are based mostly on the geometry of the edifice and outer slopes of the deposits. In general, tuff cones and tuff rings are constructional landforms where tuff cones have high relief with steep outer slopes, and tuff rings have moderate relief and moderate-to-low-angle outer slopes. In both tuff cones and rings, the bottom of the vent crater is above the overall ground surface. Maars (now typically referred to as maar-diatremes; White and Ross, 2011) have low-angle outer slopes, and the crater excavated the pre-existing landscape and rocks. Dohrenwend et al. (1986) did not measure the vents they referred to as tuff rings, but V13 (which is filled with a scoria cone) is the best exposed with cone height ~45 m, width ~1200 m, and outer slope angles ~10° (Fig. 2-1).
Geologic mapping, geochronology, and paleomagnetic studies at Cima have been instrumental in refining the techniques and resolution of these methods and is one of the locations where the terms monogenetic, polygenetic, and polycyclic have been used to understand the timing of volcanic activity (Turrin et al., 1984; Renault and Wells, 1990; Wells et al., 1994; Wilshire et al. 2002a-d). Monogenetic eruptions and volcanoes are episodic, small, and short-lived. Polygenetic basaltic eruptions and volcanoes have eruptive episodes occurring repeatedly from the same vent for long periods of time, form large volcanoes with long-lived magma chambers, and can have large changes in the compositions of erupted magma. The term "polycyclic" was suggested for scoria cone volcanoes that show evidence of having had long time gaps (thousands to tens of thousands of years or more) between eruptions, but Valentine et al. (2021) caution using this term because it can imply long-lived magma reservoirs similar to polygenetic volcanoes. Polygenetic and polycyclic terms and models were developed to address complicated data, and they might be applicable for specific volcanoes and volcanic fields, but in a recent review of monogenetic volcanism, Valentine et al. (2021) describe complicated monogenetic volcanoes. Valentine et al. (2021) suggest it is in part how one defines vents compared to volcanoes and fields and the importance of the history of a single vent. Monogenetic volcanoes can be short lived, erupt different compositions, have different styles of eruptions, have quiescent cycles, and in a volcanic field a later monogenetic volcano can overlap or be colocated.

The Cima field has been the focus of petrologic studies of volcanic rocks including magma source, potential crustal contamination, the petrology and source of xenoliths, and variations and structure in the mantle. Volcanic petrology, magma source, and possible crystal contamination have been described by Farmer et al. (1995a, 1995b), Glazner et al. (1991), Glazner and Farmer (1992), and Wilshire and McGuire (1996). Xenolith petrology and sources have been described by Wilshire et al. (1988), Wilshire (1990), Wilshire et al. (1991), Wilshire and Musaka (1997), Lufﬁ et al. (2009), and Bernard and Behr (2017). These papers show that what was partially melted in the mantle and where it was partially melted (different pressure, temperature, and depth) cause variability in the magma, and that some of the magma interacted with crustal rocks. The xenolith studies indicate heterogeneity in the mantle and that melts were injected into the mantle rocks, but that another magma included the xenoliths to bring them to the surface. Some studies suggest that the compositions of xenoliths indicate stratified mantle and that melting occurs at shallower depths with time, and Lufﬁ et al. (2009) suggest that sources resulted from tectonically subcreted and imbricated lithosphere having an oceanic protolith.

A number of surficial processes studies have investigated lava flow emplacement, erosion, and pedogenic (soil forming) processes in the Cima field. Soldati et al. (2017) used the youngest lava flow to model emplacement dynamics for rheology and morphology. Surface modification of scoria cones and lava flows have been described by Wells et al. (1984), Dohrenwend et al. (1984), and Dohrenwend et al. (1986). The rates of pedogenic processes in the Cima field have been described by McFadden et al. (1984).

The Cima volcanic field has been the target of early remote sensing studies and evaluations of methods. Early radar explored X-band and L-band wavelengths and images, including LVH, LHH, XHH polarization. In L-band LHV images the youngest lava flow was very well imaged because the cross-polarized returns had multiple scatterings from very rough surfaces of the lava flow. The older flows were weathered and eroded and had relatively low cross-polarized returns. Alluvium and desert pavement areas were very smooth and therefore dark on both LHV and LHH polarizations (Elachi et al., 1980). Landsat Thematic Mapper has been used to characterize lava-flow degradation (Arvidson et al., 1993).

The Cima volcanic field has also been used for comparisons in evaluating the volcanism at Yucca Mountain, a site studied to host a repository for high-level radioactive waste in the United States (Wells et al., 1990; Wells et al., 1991).

* D2 S1-AS1 Alternate stop at Aiken Mine with scoria cone, mantle xenoliths, and lava flows
  * 9.3 mi. UTM 11S 615914 3899494
  * Distance is total distance from D2 S1-1. See Gans (2022) for detailed descriptions.

* D2 S1-AS2 Alternate stop at V13 with tuff ring, scoria cones, lava flows, and lava tube
  * 4.6 mi. UTM 11S 610312 3896587
  * Distance is total distance from DS22 D2 S1-1. V13 is the best-preserved tuff ring in the Cima basalt field, and deposits are exposed along the road to Aiken Mine. See Gans (2022) for detailed descriptions.

D2 RL01 D V8 scoria cone and lava flows
  0.7 mi. UTM 11S 612592 3896092

The V8 scoria cone and xenolith-bearing lava flows have been extensively studied. They are located is 0.7 NE of Stop D2 S1-1. Geologic mapping, petrography, and geochemical and isotopic variations are described for the V8 cone and lavas (Waits, 1995). Waits (1995) concluded that (1) the peridotite-bearing lavas did not undergo cryptic crustal contamination, and contamination might be limited to nonxenolithic lavas, (2) there is no geochemical and isotopic variation at individual cones, (3) xenolithic-bearing lavas do have crustal material and interaction, and (4) low Mg#s indicate fractionation of olivine deep in the mantle before entraining mantle xenoliths.
Depart Stop D2 S2-1, turn RIGHT, and proceed to Kelbaker Rd.

D2 RL01 D Scoria cones V5, V6A and V7, Tuff ring V6B, and Flow QTf6B-1

0.3 mi. UTM 11S 612239 3894586

1:30 to 4:40 oc – To the west of the road, along the skyline, three scoria cones are aligned NNE, and these are identified as V5, V6A, and V7 by Wilshire et al. (2002) (Fig. 2-2). V7 is 560 ka (Wells et al. 1984, vent E). Dohrenwend et al. (1986) included V7 in the compilation of morphologic measurements of scoria cones, and V7 is near the mean or above the mean for the cone height, crater width, cone width, outer slope angle, height to width ratio, and crater width to cone width ratio. V5 and V6A were not measured by Dohrenwend et al. (1986), but topographic profiles of these two vents are consistent with those of scoria cones, and the three cones have slight differences in the amounts of erosion and rills (mapping by D.C. Buesch). Thus, V5, V6A, and V7 represent explosive magmatic eruptive events.

To the west, about 15 m above road elevation is a low, flat, arcuate ridge that is the edge of a lava pond in V6B. The lava spilled out from the pond to the ESE, and formed a lava flow (QTf6B-1) that traveled ~5.5 km to the SW and W. The lava flow ponded in V6B delineates ~135° perimeter of an ~600 m diameter crater (mapping by D.C. Buesch). Most of the V6B crater rim has eroded, and only ~60° is exposed in the south, but the diameter of the rim is ~750 m. The height of the rim is difficult to determine because the lower slopes of the vent are eroded or covered by younger deposits, but only 18 m of the outer slope is exposed with a slope of 12°. These morphometrics indicate that V6B is a tuff ring. Exposures along this rim are ash to lapilli tuff with blocks, many of granitoid rocks up to 1 m diameter, consistent with being a tuff ring that is herein identified as V6Btr. Southwest of V6A is another 80° arcuate ridge that also contains ash to lapilli tuff with blocks, many of granitoid rocks up to 1 m diameter. This ridge is a rim with minimum interpreted height of 21 m, an outer slope of ~9°, and a crater diameter of 760 m that is herein identified as V6atr. Southwest of V5 is a circular lava pond with a diameter of 125 m and no exposed rim that is herein identified as V5Btr. The three tuff cones indicate an early phreatomagmatic eruption history that has not previously been recognized, and reinforces the geometry of overlapping of tuff rings as well as potential for approximate colocation of phreatomagmatic and magmatic vents.

Additions descriptions of the V5, V6A, and V7 scoria cones, and V5Btr, V6atr, and V6Btr tuff cones (with different names) are in Gans (2022) Stop 4.

D2 RL02 Kelbaker Road – Turn LEFT

2.8 mi. UTM 11S 610167 3890932

D2 RL03 Kelbaker Road & Black Tank wash Rd – Turn RIGHT

4.3 mi. UTM 11S 603977 3893957

D2 RL05 D Youngest lava flow in the Cima volcanic field (QTf2B-1)

0.8 mi. UTM 11S 605023 3894240

The low-lying, low relief exposures of basalt to the north of the road have assigned ages of 130 ± 30 ka and 90 ± 70 Ma (Wilshire et al., 2002). On the right of the road next to the basalt flow is the distal end of 'a‘ā flow QTf2-1 (QTf2B-1) from V2B (or Qv3lb from Qv3 in Wells et al. 1994). This is the youngest lava flow in the Cima volcanic field with a \(^{3}He\) age of 13 ± 3 ka from the distal end of the flow (Wells, 1995), and an average \(^{36}Cl\) age of 11.5 ± 1.5 ka from two samples from the same flow (Phillips, 2003). The top of the flow is very rough and littered with clinker from the lava and reddish scoria, some of which have fusiform shapes, that are probably rafted pieces from the vent. The lateral margins have numerous small lobes and embayments that resulted from small break outs. This youngest lava flow has been modeled for emplacement dynamics for rheology and morphology (Soldati et al., 2017).
STOP D2 S1-2 Black Tank wash with interstratified lava flows and V3 tephra deposits

0.7 mi. UTM 11S 607000 3894370

This stop is in a wash that separates the V3 scoria cone source for the lava flow QTf2-1 from vent V2B. V3 cone has a breach on the western side down which lava flowed, and this lava has an assigned age of 0.58 ±0.16 Ma (Wilshire et al., 2002). The wash slopes at a low angle to the west, and the stratigraphic section slopes slightly more steeply than the stream, so a walk along the wash to the east exposes in stream banks rocks progressively deeper in the section. More detailed descriptions and interpretations are in Gans (2022, Stop 2B).

The top of the section is the 'a'ā flow QTf2-1, which has been exposed from the western end of the wash. A short walk to the top of the flow is through a thick upper rugged block-and-clinker zone that has 1-6 meter high 25-35 m wavelength arcuate ridges that delineate flow directions and channels. Locally on the lava flow are concentrations of scoria that are pieces of the source cone that have been rafted by the lava flow. The thick middle of the flow is sparsely vesicular and locally there are well developed, but contorted, flow bands. Cooling of the flow created well-developed cooling fractures and joint-bounded columns that do not match in the middle of the flow where there is a subhorizontal cooling parting. This mismatch of columnar cooling fractures indicates that fractures propagated from the top and bottom toward the middle of the flow as it cooled. Locally, there are numerous paleomagnetic core drill holes. The base of the flow has a lower clinker zone. This one could have formed in place resulting from shear developed along the cooling base, or the clinker was formed on the top of the flow, was carried to the front of the flow where it temporarily formed the flow front and was then overrun by the flow. The analogy of this motion is that of a tractor tread.

Below the 'a'ā flow is a polymeric (volcanic clast) pebbly to bouldery coarse sandstone that appears to be on the south side of the wash, but not the north side. The upper part of the sandstone is reddened and the most reddened is a 30 cm thick sandstone that is overlain by three gravelly beds and the uppermost bed is a 4 cm thick fine-grained sand (possibly aeolianite). These reddened beds are likely a “baked” zone formed from the overlying lava. Wells et al. (1994) collected a sample of the sandstone 25 cm below the base of the lava flow for thermoluminescence (TL) chronology, and the sediment was thermally reset to yield an age of 8.5 ± 0.7 ka to 9 ± 0.8 ka depending on the assumed water content of the sandstone (S. Forman, pers. comm., 1992) in Wells et al. (1994). The amount of reddening in these five beds might be a result of how much moisture was in each bed.

To the east in this wash is a 1-4 m thick 'a'ā flow that forms a dry waterfall where it is overlain by the sandstone. To the west of the dry waterfall, the lava thins and pinches out as an apparent flow lobe terminus. To the northeast of the dry waterfall, the lava flow thins, and might indicate an onlapping of the flow with the southern edge of V3.

The lava flow overlies a gray to yellowish brown, well-bedded, lithic-rich lapilli tuff. Clasts are angular basalt, highly vesicular light-brown small vesiculated fragments, and granitoid clasts. Beds are 1-30 cm thick, and bedding structures indicate lateral flow by pyroclastic density currents (surge). On the south side of the wash, these beds have blocks up to 60 cm diameter that form impact sags and represent ballistic impacts. These beds have low-angle dips to the south and are interpreted as outflow sheets from early phreatomagmatic eruptions that formed a tuff ring, largely obscured by the V3 scoria cone.

Farther up wash to the east, there is another dry waterfall formed by a lava flow that is interstratified with sedimentary rocks that lie above the western dry waterfall and the QTf2-1 flow. This wash exposes a complex interstratification of three lava flows and three types of sandstone, tuffaceous sandstone, and lithic lapilli tuffs that formed either as part of the early V3 tuff ring or were deposited against the tuff ring.

* D2 S1-AS3. Alternate stop at V12 with tuff cone, scoria cones, and lava flows

* 4.6 mi. UTM 11S 610312 3896587
* Distance is total distance from D2 S1-2. V12 has very good exposures of tuff ring deposits along the south side of the drainage. See Gans (2022) for detailed descriptions for his Stop 3.

Depart D2 S1-2 and proceed to Kelbaker Rd.

D2 RL04 Kelbaker Road – Turn RIGHT

1.5 mi. UTM 11S 603977 3893957

STOP D2 S1-3 17 Mile Point – Lava flow and pre-lava rocks, geomorphology, and desert pavement

1.5 mi. UTM 11S 602652 3895821

Turn LEFT onto Mojave Road (dirt road), and park in the open area near the highway berm.

In the wash on the east side of the road, a 4-6 m thick basaltic lava flow was deposited on well-stratified arkosic (grus) alluvial or fluvial sandstone and conglomerate. The upper 1-2 m of the gravelly sandstone is slightly reddened, representing a baked zone or a poorly developed paleosol. The base of the lava is vesicular clinker and a sheared lava subzone, and oblate vesicles near the base of the flow are slightly imbricated to the base indicating flow to the west. This flow direction is not a surprise, but it simply shows that oblate and imbricate vesicles can indicate direction of flow. The sparsely vesicular core (subzone) of the flow has some faint textural banding. Up section, the vesicles increase in size and abundance in the upper moderately vesicular subzone, and locally the foliation of oblate vesicles indicates fold deformation. The upper 1-2 m has an increase in sheared fragments that represents the lower part of the upper vesicular sheared and clinker subzone.
of the cooled top of the flow. The fracture characteristic (orientation, trace length, planarity (>1 m length) and surface roughness (<1 cm) differs in each subzone. During the climb to the SW end of the flow, notice the flow was deposited on Proterozoic gneiss that appears to have formed a bedrock hill that merged into the hill to the west of the road. Except for this local occurrence of lava flow deposited on gneiss, sandstone deposits on metamorphic rock occur on both sides of the road, and on the west side the sandstone banks against the metamorphic rock. Evidently, the wash passed north of this hill of bedrock prior to the emplacement of the lava flow. The wash deposit is at a lower elevation than the bedrock and cannot have graded to cross the bedrock in the vicinity of the highway.

On the top of the mesa-forming flow, the surface is vastly different from the top of the young flow at the D2 S1-2 stop we just visited: it is somewhat rolling but fairly smooth, and largely covered by a desert pavement composed of lava fragments. Note that (1) there are no other rock types in the pavement, (2) the fragments are not rounded, (3) stream geomorphology is not present. These factors preclude a stream flowing on the crest of the lava flow and force rethinking of the long-held origin of desert pavements as lag surfaces resulting from stream flow. Furthermore, if we were to dig a pit here, we would find only silt-size material, all pale brown in color. A chemical analysis indicated that it is similar in composition to the Soda Lake playa deposits and distinctly unlike that of basalt. This is reconciled by the postulate that the silt cap is eolian, and the desert pavement must rise with the addition of silt. Thus, was reasoned the “born on the surface” model for desert pavements (Wells et al., 1995), with the eolian origin of the vesicular A horizon (McFadden et al., 1987) and the fact that Mojave Desert soils are cumulic; soils formed by accumulating matter, in this case dust (McFadden et al., 1987).

An interesting twist to the landform modification story is displayed here. A careful examination of this surface will reveal sparse large crystals of potassium feldspar. Upslope on this surface, none of the clasts are present. It is probable that the basalt flow dammed the drainage that lay along the north slope of the old bedrock, the stream infilled against the bedrock and basalt, and it eventually overtopped the low point in the basalt. Support for this hypothesis is present to the west where large fragments of basalt lie midway up slope on bedrock. The stream overtopped its basalt dam and cut down through basalt as well as gneiss to establish a new stream course that is lower than the original.

**Lunch** – We have about 20-30 mins for lunch.

_Depart Stop D2 S1-3 and turn LEFT on Kelbaker Rd. toward Baker (~14 miles)._ 

**D2 RL06 D Lava and pre-lava sedimentary rocks**

0.3 mi. UTM 11S 602678 3896278

This location is at the north end of the lava flow examined at S2-3. The mesa capping lava flow (from D2 S1-3) with an assigned age of 0.58 ± 0.16 Ma overlies thin bedded gravelly arkose (composed of grus) that is similar to the sandstone and conglomerate in the wash south of the lava flow at the last stop.

Where the road rises across a basalt at the NW end of the mesa, this basalt has an assigned age of 0.17 ± 0.06 Ma (Wilshire et al., 2002c). The base of the flow is locally exposed, and it was deposited on well-bedded fluvial gravel. This and the narrow width and long runout traced back to the east indicates the flow was confined to a wash that was similar to the modern wash. This basalt compared to the mesa capping basalt is a good example of an inverted topography.

**D2 RL07 D Dirt road turnoff to Indian Springs**

0.8 mi. UTM 11S 602256 3897472

* D2 S1-AS4. Alternate stop at Indian Springs with lava flow deposited on banded gneiss

* 3.9 mi. UTM 11S 607887 3899537

* The Indian Springs dirt road is bounded on both sides by a Wilderness Area, so please stay on the road. The road ends in 2.9 miles at a large sandy wash. Park, and walk ~1.0 miles to the lava flow. See Gans (2022) for detailed descriptions for his Stop 7.

**Continue to Baker**, passing inselbergs of Cretaceous granite along the way. The broad piedmont has scoria clasts that under study by Andeski and Hughson (2022).

**D2 RL08 Baker & I-15 – I-15 west-bound freeway on ramp.**

TURN LEFT crossing the channel connecting Soda Lake to the south and Silver Lake north of here. This channel flows during Mojave River flood events.

14.0 mi. UTM 11S 584200 3902780

**D2 RL09 I-15 & Afton Rd (AC9614) – Off-ramp**

25.2 mi. UTM 11S 553470 3881213

Possible assembly area in large dirt area north of I-15 to collect caravan stragglers. We are parked on the youngest highstand beach of Lake Manix.

**D2 RL10 Afton Rd (Powerline Rd) (CL8337)**

End of pavement, having passed exposures of Lake Manix sediment including tufa-lined layers.

1.3 mi. UTM 11S 551927 3880004

**D2 RL11 Playa 1 on right**

1.8 mi. UTM 11S 550347 3882740

**D2 RL12 Playa 2 on right**

1.0 mi. UTM 11S 550005 3883823
D2 RL13 Playa 3 on right.
This playa supports large creosote bush and crucifixion thorn, both unusual for playas and the latter is more widely occurring in Arizona.

0.7 mi. UTM 11S 549744 3884840

D2 RL14 Afton Rd-Powerline Rd (CL8315) – Turn RIGHT
2.4 mi. UTM 11S 548479 3887629

D2 RL15 D Pliocene sandstone, SE tilted with small faults
0.7 mi. UTM 11S 549094 3888489

D2 RL17 Y junction – BEAR LEFT
0.8 mi. UTM 11S 549864 3889539

D2 RL18 Y junction with BLM 3813 – BEAR LEFT
0.1 mi. UTM 11S 549944 3889674

STOP D2 S2 -Powerline sequence and Bicycle Lake basalt

STOP D2 S2-1. Overview, avalanche breccia, and Powerline sequence tephrochronology
0.2 mi. UTM 11S 549997 3889898

Park along road.

Stops for this afternoon will be outside the southeast boundary of Fort Irwin, where we will examine the southernmost exposures Bicycle Lake basalt and the Powerline sequence of groundwater discharge deposits (GWD) and playa deposits upon which the basalt was deposited. The basalt is exposed on the Whale, Squid, and a small exposure south of the Squid (Fig. 2-3). The Whale was named for the ridge in the mid 1800s along the Old Spanish Trail where Bitter Springs was an important watering stop (Lyman, 2004). Mapping of these rocks supports the groundwater resource management at Fort Irwin and the evaluation of faulting in the ECSZ.

Although this Desert Symposium field trip is focused on young basalts, those basalts also have interesting implications for surrounding rocks in many cases. We will explore one such example in this area. Where we have stopped we can examine the transition from fluvial sedimentation to the west (Byers, 1960) to wetland sedimentation to the east. The dividing point is approximately at a catastrophically deposited rock avalanche (Fig. 2-4). More broadly, we can follow the wetland sedimentary rocks northeastward to where they are overlapped by the Bicycle Lake basalt, and yet farther to where the basalt lies on playa deposits. There evidently is a gradation in depositional environment from fluvial to wetland to playa.

Stop 2-1a Avalanche breccia
The rock avalanche deposit is exposed in the ridge a few tens of meters southwest of this location, where it lies on northeast-dipping fluvial gravels that overlie rocks of the Barstow Formation. The Barstow Formation near here contains a tuff bed that was dated by Swisher (1992) at ~12.5 Ma. The rock avalanche is composed of several types of plutonic rocks, ranging from porphyritic granite gneiss to diorite. Aplite dikes are notable in several places. The breccia is persistently fractured and yet dikes are essentially continuous, with little disruption. This jigsaw brecciation with little matrix material is
characteristic of rock avalanche deposits in which giant slabs of bedrock are thoroughly broken as they slide down a steep slope and over gentler landscape, but they are not mixed and jumbled (Bishop, 2013). Spectacular exposures across the powerline road to the south display features such as jigsaw fracturing, sandstone dikes, and folded underlying strata. These features point to a rock avalanche origin for the breccia, and the rock types correspond generally to those seen in mountains to the south near and in Cave Mountain. Huge infolds of sedimentary rock are present in several places within the avalanche breccia, which may indicate that several successive avalanche sheets were emplaced, the younger ones deforming the older. Alternatively, the avalanches may have entrained sediment.

Where was the source of the avalanche? This seemingly simple question is made difficult by the old age of the avalanche. The basin sediments lying on the avalanche are as old as ~9 Ma, so modern mountains are probably quite different from those that provided the relief and gravitational potential energy for the avalanches. However, we can try to match lithology with exposed rocks: Alvord Mountain to the west fails because of the lack of clasts of Paleozoic rocks, Cretaceous dike rocks, and the widespread basalt of that area (Byers, 1960). The Tiefort Mountains area to the north fails for a reduced amount of diorite there and generally weakly foliated rock; however, parts of the area might be a suitable source (Schurme et al., 1996). The Soda Mountains to the east have much metavolcanic rock and the mountain is topped by a volcanic stack, so it also fails as a source. The Cronese Hills on the south have distinctive metasedimentary rocks that are not present in the breccia, so it also fails. However, Cave Mountain (~15 km SE of this location), and mountains immediately to the north, may be the source. Both lithologic types of the breccia are present and much strongly foliated rock is also present. In addition, thick avalanche breccia lies near Cave Mountain and young landslide or avalanche masses lie on the south side of the mountain. This source inference leads to an important conclusion: that the Cave Mountain area has been topographically high at least two times, and perhaps for an extended period of greater than 10 million years.

At this location, deposits overlying the rock avalanche are thin-bedded calcareous sandstone and siltstone, typically brown in color and in many places cemented into a porous sandy limestone or nodular ‘popcorn’ rock: a groundwater-discharge deposit (GWD). One ridge within this sedimentary section is anomalous, consisting of tilted beds of similar rocks (but with chunks of avalanche material) as well as large rip-up blocks, and lenses and sheets of breccia derived from leucogranite. Along strike

Figure 2-4. Geologic map of the powerline sequence area. Tbl, Bicycle Lake basalt; Tps, Powerline sequence that includes Tpbg, a gravel unit; Tav, avalanche breccia; Tsb, sedimentary breccia; Tbg, fluvial gravel. All units with “Q” are Quaternary in age. Heavy black lines are faults (dotted where concealed); heavy blue dashed lines represent opal beds; light blue dashed lines are marker beds. Strike and dip symbols in black show bedding orientation; dip given in label. Note the interlayered geometry of gravel and avalanche breccia. Mapping by D.M. Miller and D.C. Buesch.
within this sequence a rock avalanche thickens to the north, demonstrating that avalanches occurred after the transition to wetland deposits (Fig. 2-4).

Upward in the section, wispy opal beds, abundant rhizoliths (here formed by opal), and high content of calcium carbonate attest to a wetland (groundwater discharge deposit) origin of the sediment. Walk east along the south side of outcrops to examine the sediments. Upward, a prominent white bed of ash as thick as 25 cm is seen, the lowest of several in the sedimentary sequence. Above that ash is a ridge-forming white opal bed that can be followed for considerable distance near the base of the wetland sequence (dashed blue line in Fig 2-4).

The sequence we see here above the basal rock avalanche is characteristic of a small, but unusual, sedimentary sequence that we refer to as the Powerline sequence. It is unusual in being almost entirely deposited under wetland conditions, that is, shallow groundwater, and with subordinate stream deposits, which is distinctly different from the widespread fluvial and alluvial fan facies for young deposits across a wide expanse from Alvord Mountain to the Cronese lakes. Within this basin are numerous ash beds, some of which have preliminary chemical correlations with dated ashes elsewhere (Walkup et al., 2022).

Stop 2-1b Tephra stratigraphy and chronology

In the Powerline sequence, at least 12 tephra deposits have been mapped (Fig. 2-5). The lowest tephra layer (sample LCW-MOJ21-001) is exposed below the opal deposits. It is 24 cm thick and white to light gray in color. The glass shard composition correlates well with either the ca. 9.3 Ma McGuire Peak ash bed (Snake River Plain source) or 8.9-8.6 MOD-3/19 ash bed (proto-Cascade source) in the marine Modelo Formation near Ventura, CA (Walkup et al., 2022). Until we get more information allowing us to determine which correlation is more robust, we can at least say that this lower tephra layer is age constrained to 9.3-8.6 Ma.

In this same low-relief area of the basin, continuing to the north-northeast and moving up-section, there is a dark gray dacitic tephra (LCW-MOJ21-013) and another white tephra (LCW-MOJ21-015). The dacitic tephra has been analyzed but does not have any close matches to tephra with known ages. LCW-MOJ21-015 has not yet been analyzed.

To the north in the cliffs on the west side of the powerlines, another eight tephra layers are exposed, one of which may be the same tephra layer as LCW-MOJ21-015, and the remaining five tephra sites (some of which are visible at Stop 2-2) are found to the northeast on the east side of the powerlines. A closer view of this stratigraphy is visible from where we turn down the broad central wash.

Return to vehicles and proceed northeast to rejoin the main powerline road.

Figure 2-5. Generalized composite tephrostratigraphic section of Powerline sequence tephra layers (Walkup et al., 2002). Section has not been measured; thus, the tephra positioning is relative. The relation between M16SM-3299 and the rest of the tephra sequence is unclear and the possibility of fault offset is high, so its vertical position relative to the rest of the section is uncertain.

D2 RL19 Turn RIGHT down a broad central wash.
Ahead is a basalt-rimmed bluff and ash beds are exposed on both sides of the wash as we proceed downstream.

0.8 mi. UTM 11S 550667 3890712

STOP D2 S2-2 Bicycle Lake basalt

0.5 mi. UTM 11S 550890 3891030

Along the north side of the broad wash exposures of the Powerline sequence include the uppermost tephra deposit, the base and lowermost flows of the Bicycle Lake basalt, and a series of faults related to the sinistral Coyote Lake fault (Fig. 2-3, 2-4). The south side of the wash has another tephra from the Heise volcanic field, Idaho.

The uppermost tephra layer (DM-BS-1) is exposed just above stream grade. It is a white to light gray tephra that here is offset by several small faults. The glass shard composition of this sample correlates well with several tephra collected from the Darwin and Coso Wash areas in California and is likely of Coso volcanic field origin, providing a broad age constraint of 5.8-3.0 Ma (Walkup et al., 2022).
et al., 2022). The Bicycle Lake basalt caps this section. Different geochronologic methods have yielded ages ranging 5.6-2.9 Ma for the basalt, and a new $^{40}\text{Ar}/^{39}\text{Ar}$ eruption age of 4.55 ± 0.07 Ma has been determined on a basalt flow from the southwestern part of the Whale (Buesch et al., 2022).

Due south across the wash there is a half-meter thick exposure of tephra (sample LCW-MOJ21-004). This sample correlates well with several tephra derived from the Heise volcanic field in Idaho and provides an age constraint of 7.1–6.27 Ma for this tephra layer (Walkup et al., 2022). This same layer also occurs near the base of the cliff directly west of this stop.

A series of closely spaced, 070° striking faults project to the east through the Powerline sequence GWD and Bicycle Lake basalt. These fault spays are consistent with projected alignments of the Coyote Lake fault. A set of small cross-faults that is probably part of the detailed fault network is exposed near where the basalt approaches wash grade. At this location, the faults appear to disrupt the northern edge of a broad low-relief channel incised into the pre-basalt sedimentary rocks.

The pre-basalt sedimentary rocks include the Powerline GWD that were locally eroded to form low-relief gullies and broad swales filled with polymictic sandstone and conglomerate. The largest clast at the top of the conglomerate is 10x55 cm, red and medium gray, coarse-grained granitoid. The clast population of plutonic and metamorphic rocks in these younger deposits is similar to that in some of the interstratified sandstone and conglomerate in the GWD, and it probably had the same provenance. Some of these clasts might be locally derived (recycled) from the erosion of the GWD, and clasts of carbonated-cemented sandstone are in the conglomerate. Locally, the sandstone and conglomerate below the basalt is light yellowish gray and not oxidized; however, there are also locations where the sandstone and conglomerate are light reddish brown. Lava flows (with temperatures typically 800-1,200°C) can “bake” or oxidize and partially fuse subjacent sediment. Baking occurs because the temperature of the lava radiates down into the substrate, and if the sediment is moist, the water turns to steam, oxidizing the available iron in the sediment. If there is no moisture, there is no steam, and thus no oxidation or visible baking. Fusing of sediment can occur if there are clays in the sediment. The lack of oxidation in the pre-lava sediment indicates that these deposits were not “baked” by the lava and probably represent dry deposits. The light reddish brown sandstone and conglomerate probably has this color resulting from paleosol development, indicating that these deposits were at a stable geomorphic surface. The importance of this location is that it shows (1) the GWD and wetland environment existed for a long period of time, (2) the system dried out, possibly related to slight uplift or lowering of local base level, (3) a broad low-relief geomorphic surface developed an incipient paleosol on the interfluvus with development of low-relief (<1-2 m deep) gullies and swales, (4) somewhere nearby, enough relief was generated to incise into the GWD deposits and recycle clasts and transport them to this location, and (5) the Bicycle Lake basalt field formed and distributed lava flows to this area.

The Bicycle Lake basalt was deposited on the Powerline sequence, and these exposures of the basalt are among the southernmost in the field. The lava flows are typically 1-2 m thick, but locally they are up to 4 m thick and are interstratified with thin lava flow cosets up to 4 m thick. [Paraphrasing from Fisher and Schmincke (1984) and Jackson (1997), a co-set (or coset) is a set or sequence of beds that have similar characteristics such as texture, structure, or composition that set them apart from other beds above or below. Cosets can be applied to sequences of similar lava flows (Buesch et al., 2022).] Lava flows have typical textural and structural zoning based on three main textures (1) chilled margins, (2) moderately vesicular, and (3) sparsely vesicular, and these are vertically symmetrical around the core. The core typically is sparsely vesicular (< 5 percent vesicles >2 mm); above and below the core are the upper and lower moderately vesicular subzones (typically 20-35 percent vesicles >2 mm), and at the top and bottom of the flow are the upper and lower chilled margins that are sparsely to slightly vesicular subzones. Some flows develop a thin (<20 cm thick) lower prolate vesicle subzone above the chilled margin. Some flows have prolate vesicles and vesicular pipes (both of which typically developed as vertical textures or structures because of the buoyant rise of the vesicles), and some flows have vesicular sills. Some flows have imbricate oblate vesicles in the lower chilled and lower moderately vesicular subzones, and (or) bent prolate vesicles that can be used to determine flow directions. Flow directions in the Squid typically indicate flow to the SSE-to-SSW, but some flows in the southern Squid flowed SW-to-WSW. On the edge of the ridge to the west of the powerlines is an 80-cm thick flow (but only the sparsely vesicular core to the base is exposed, and the upper moderately vesicular has been eroded or never formed). In addition to a flow direction to the SW, this west of the powerlines flow is the most distal southwestern flow on the edge of the volcanic field.

**Return to the powerline road, turn RIGHT.**

During the drive up the hill, eight tephra layers (LCW-MOJ21-034, LCW-MOJ21-033, LCW-MOJ21-010, LCW-MOJ21-005, LCW-MOJ21-006, LCW-MOJ21-003, Heise-derived tephra LCW-MOJ21-008, and LCW-MOJ21-009) are exposed to the west of the road, and two tephra layers (LCW-MOJ21-009 and DM-BS-1) are exposed on the east side of the road.

**STOP D2 S2-3. Overview of the south end of the Whale (from the top of the Squid).**

0.9 mi. UTM 11S 551747 3892352
From this flat ridge top on the north side of the Squid there is a panorama of the Cronese basin that extends along the northwestern to southeastern skyline. This basin includes the drainages in the southeastern Fort Irwin, the U.S. Army National Training Center, and continues to the Cronese Lakes near I-15. This is one of the 13 groundwater basins in Fort Irwin.

<table>
<thead>
<tr>
<th>Direction*</th>
<th>Feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>12:00</td>
<td>Reference direction along the NE-trending road.</td>
</tr>
<tr>
<td>11:55</td>
<td>Stop 4 is near the southeast base of the Whale near the power lines.</td>
</tr>
<tr>
<td>2:30-to-11:25</td>
<td>Soda Mountains (skyline) to Red Pass ridge (intermediate distance) are the eastern groundwater basin divide (with small internal basins).</td>
</tr>
<tr>
<td>12:10-to-11:30</td>
<td>Moderate relief hills of avalanche breccia in the topographic basin east of the Whale.</td>
</tr>
<tr>
<td>11:55-to-9:30</td>
<td>The visible western edge of the Whale is ~9:36. The Whale with SW dipping Bicycle Lake basalt has numerous kink and faulted folds highlighted by the light gray sand decoration. Along the entire western side of the Whale are dip slopes on the basalt that dip 25-40° SW. Higher on the slopes are areas that have typical dips of ~3° and ~10°.</td>
</tr>
<tr>
<td>9:30-to-7:10</td>
<td>At the northwest end of the Whale, one of the Spanish Trail routes used during the 1850s came from Bitter Springs, rounded the end of “the whale”, and headed ~15 miles up the alluvial washes to the Impassible Pass near Spanish Canyon at Alvord Mountain. From this vantage point, the Pass is ~S70W, just behind the far distance (skyline), low-relief, triangular profile, dark brown hill. Tiefort Mountain (the highest skyline point). Bicycle Lake and Bicycle Lake mesa are at the west end Tiefort Mountain, but they are obscured by the intermediate distance ridge. North of the intermediate distance ridge is the “Valley of Death” (named by the military), and it follows the Bicycle Lake fault, which also forms the steep-sided north end of the Whale. About 1.8 km away is a dark brown hill at the northwest end of the Squid, and this is pyramidal (three-panel) dome of basalt with underlying Powerline GWD deposits in the core that forms the highest point of the Squid.</td>
</tr>
</tbody>
</table>

* Direction X-to-Y is a counterclockwise sweep.

![Geologic map of the southern Whale and Squid area](image_url)

Figure 2-6. Geologic map of the southern Whale and Squid area. Arrows point to fold axes in basalt that show in the hillshade base map.
The geologic map of the southern Whale and Squid shows the distribution of the basalt, the pre-basalt Powerline sequence and playa deposits, folds and faults in the Whale and Squid, and the projection of the Coyote Lake fault between the two (Fig. 2-6). On the map, the Powerline sequence and the playa deposits are overlain by the Bicycle Lake basalt, but detailed stratigraphic relations are not yet known.

This is a good place to appreciate a major change in depositional environment. To the southwest, at stop S2-1 we stood at deposits representing distal fluvial environments that were capped by rock avalanche and gave way upward to wetland environments. To the north, the brownish badlands below the Bicycle Lake basalt are playa mudstones, several tens of meters thick. The lower part of the playa deposits probably represents the depocenter for the streams that led from the west to the playa. We postulate that the arrival of a lobe of rock avalanche (with its positive topography) diverted streams, but groundwater levels remained shallow in the area between the avalanche lobe and the playa, driving wetland sedimentation of the Powerline basin.

At the Whale and Squid, comparing the physical stratigraphy (number of thick flows, flow thickness, interstratification of thicker flows with thinner flow cosets, flow directions) and geochemistry indicates similarities, but some differences, in these two areas (Buesch et al., 2022). The typical stratigraphic sequences of the Squid have fewer flows, fewer interstratified flows and cosets, and flows thin southward. Typical flow directions in the southern Whale are to the SSE-to-ESE with most to the SE, and at the Squid they indicate typical flow directions to the SSE-to-SSW. Geochemically the flows at the southern Whale and Squid have the same range in compositions. The physical stratigraphy and geochemistry are compatible with the southern Whale and Squid being part of the southern end of the Bicycle Lake volcanic field, and the Squid being near the margin. The flow directions are compatible with a general flow to the SE, and the distal flows (represented in the Squid) might have responded to local paleotopographic redirection, but this might be at odds with some of the inferred flow directions of the pre-basalt lower playa deposits. Alternatively, the Whale and Squid might represent different parts of the southern basalt field that have been juxtaposed by the Coyote Lake fault. At this time, these types of reconstructions based on paleotopographic influences to flow directions versus fault separation are still being considered (Buesch et al., 2022). Regardless of the final interpretation, it seems clear that the basalt flowed over a very low relief surface of wetland and playa deposits.

The morphology of the Whale is that of a W-SSW dip slope (with W at the north end and SSW at the south end) that is bounded along the western side by a narrow moderately (locally steeply) dipping structural panel that resulted from a faulted fold (Buesch et al., 2022). The flow tops near the ridge crest, especially in the southern part of the Whale, have intermittent long and narrow subhorizontal benches or fold trains (enhanced by locally deposited aeolian sand) that appear to be small versions of the faulted folds along the west-southwest side of the Whale. These folds and faults indicate W-SW compression resulting in (1) transpressional strain transferred from the Coyote Lake to Bicycle Lake faults, and (2) possibly curving of fold axes toward the south owing to drag along the Coyote Lake fault to the south (Buesch et al., 2022).

D2 RL20 Powerline access road – Turn LEFT. Proceed N35W to wide cleared area.

1.6 mi. UTM 11S 552776 3893546
STOP D2 S2-4. View of the basalt capping playa sediment at the south end of the Whale.

0.1 mi. UTM 11S 552626 3893697

Stop in the clearing near the pipeline corridor that marks the edge of Fort Irwin property. Basalt boulders litter the surface and small cuts expose reddish brown sandy muds that we interpret as playa deposits. Look to the northwest at the reddish bluff capped by black basalt. Here we are looking parallel to the axes of folds that we saw from a distance at the previous stop. The folds are evident as steps in the elevation of the basalt from this perspective. When walking on the uppermost playa beds, the folds are expressed as steep segments of basalt alternating with gently dipping segments. The playa beds are not well exposed. Farther down section, playa sediment is well exposed as parallel bedded sandy mud with distinct sand beds and rare gray clay beds. It is not clear whether the playa beds have been folded or faulted as the basalt was folded.

It is interesting that the playa beds are directly overlain by Bicycle Lake basalt, unlike the GWD beds, which are unconformably overlain by a thin alluvial fan sequence that is in turn overlain by the basalt. This relation indicates that, while much of the GWD and playa sequences are probably time-equivalent, the uppermost playa beds probably are younger than those of the GWD sequence in Powerline basin. The playa deposits are extensive and can be tracked north to the Bicycle Lake fault and east an equal distance. The change in sediments provides a snapshot of paleogeography at the time of the initial lava flows.

End of trip. Retrace the route to I-15. Distance to Barstow is 36.5 miles.

For those interested in more examination of deposits, we offer stops on the route back to: 1) ash beds in the Powerline basin, 2) rock avalanche deposits, 3) Bicycle Lake basalt, and 4) dacite domes of uncertain age.

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† Any use of trade, product, or firm names is for descriptive purposes only and does not imply endorsement by the U.S. Government.
Monogenetic and polygenetic basaltic cinder cones and lava fields in the Cima Volcanic Field, Eastern Mojave Desert

Phillip Gans
University of California, Santa Barbara

Introduction
This field trip focuses on the southern part of the Cima Volcanic Field (CVF) and will introduce you to the landforms and deposits associated with diverse eruptive styles in this fascinating area. This volcanic field is remarkable in that it has been the site of episodic short-lived small monogenetic and polygenetic basaltic eruptions from distinct vents over the last several million years, with no obvious local structural control, temporal or spatial migration pattern, or regional tectonic explanation. The youngest eruption is about twelve thousand years old (similar to the Pisgah field), and the oldest are late Miocene (~ 7 Ma). The CVF provides a superb natural laboratory to examine rates and processes of erosional degradation for diverse rock types in a desert environment and to appreciate how the geomorphic evolution of this part of the Mojave was both influenced by and influenced the distribution of various volcanic deposits.

I have been using this area as a field exercise for my Physical Volcanology class for many years, where students use field observations and Google Earth imagery to carefully describe and map different lava flows and vent facies deposits and reconstruct the sequence and style of eruptions from different vents. There are many different vents, and all lavas in the field are rather monotonous olv (± cpx ± plag) phyric basalt, but a detailed eruptive history for the Cima field can still be pieced together using a few simple “rules”:

a) Lavas seek out topographically low ground and always flow downhill. Thus, they tend to form channel-filling “finger flows” sloping away from their vents.

b) Lavas from younger vents tend to flow around the margins of older high-standing vent edifices and thick older lava accumulations.

c) Surfaces of older lavas are more degraded than younger ones with greater vegetation and eolian sand and silt cover. The pāhoehoe or ‘ā‘ā surface of older lavas is eventually entirely lost and replaced by well-developed desert pavement.

d) Older cinder cones and tuff cones are more dissected than younger ones. For the oldest vents in the field, the original scoria cone may be entirely eroded away,
leaving only a dense plug of basalt marking the location of the central vent.

e) Younger lavas or tephra can and will occasionally rest directly on older lavas or tephra—i.e. the law of superposition. As explained in (f) below, this relation is not all that common. This criterion should only be applied when the depositional contact is directly observed, because relative elevation is a poor guide to relative age at Cima.

f) Topography is commonly “inverted”, whereby older channel-filling lavas now form elevated finger mesas, and younger lava flows sit at lower elevations. This inversion is a consequence of lavas being far more resistant to erosion than the surrounding alluvium and granite basement. Once a stream channel is armored with lava, the high standing ground on either side will erode downward more rapidly, ultimately creating new and topographically lower washes, to be filled by younger lavas.

Using these simple rules combined with examination of key contacts and outcrops, we have constructed a preliminary geologic map of the southern CVF that builds on the previous mapping by Wilshire (1992), and delineates the different lava flows and vent facies deposits associated with each vent, color coded according to their relative ages. This map will serve as the basis for discussion at several of the stops. Each of the vents in the field was originally assigned a number by Wilshire (1992) and we have followed that same numbering scheme in Figure 1 and in our descriptions of individual stops below.

Geologic background

Volcanic rocks of the CVF consist of more than 40 basalt cinder cones (~50 distinct vents) and associated lava flows that overlie variably tilted Tertiary and Quaternary sedimentary rock and older Mesozoic and Precambrian basement rocks. The volcanic field covers an area of ~200 square miles and ranges from 2,132 to 4,950 feet (650 to 1,509 m) in elevation. The southern part of the field is situated on the southwestern flank of a broad topographic feature called Cima Dome. This geomorphic pediment dome is an ancient erosional feature developed on the Mesozoic Teutonia batholith and has no direct relationship to the younger volcanic activity of the CVF. The oldest vents in the CVF tend to be in the north and the youngest are mainly in the south, but there are many exceptions. The cinder cones range from 25 to 155 m in height and from 200 to 920 m in basal diameter. The largest single lava flows cover 10s of square kilometers, but most are substantially smaller. Over 30 of these cinder cones and associated lava flows are Pleistocene in age and are located in the southern part of the volcanic field, with the rest being mainly Pliocene. The oldest flows in the field are 7.6 Ma, there is a conspicuous eruptive hiatus between 3 and 1 Ma, and the youngest flow has been estimated to be between 40,000 and 400 yr B.P. (but is probably 12,000 years old; Wells et al., 1995; Phillips, 2003). It is quite remarkable that this relatively restricted geographic area has been erupting small volumes of basalt for such a long time. The long history of eruptions and well exposed arid environment of the Cima volcanic field has made it an ideal laboratory to study rates of landscape evolution in desert environments – including rates of degradation and downcutting of pediment surfaces, and degradation of cinder cones (Dohrenwend et al., 1984, 1985). Even a novice geologist can often surmise the relative ages of different cinder cones and lava flows based on their geomorphic expression and degree of degradation.

All basalts in the field are alkalic, hypersthene- or nepheline-normative hawaiites or basanites (trachybasalts) (Wilshire, 1991). Study of Nd, Sr, and Pb isotopic compositions of hawaiites less than 1 Ma in age from the Cima volcanic field suggested that they were derived from a somewhat depleted MORB-like mantle source (Farmer and others, 1991). Many of the eruptive centers have copious amounts of mafic and ultramafic lower crustal and upper mantle xenoliths entrained in the basalt lavas or in early erupted bombs on the cinder cones. These
xenoliths attest to the primitive (mantle) origin and rapid ascent velocities for these melts. Wilshire et al. (1991) discussed the possible origins of various types of xenoliths found in the Cima volcanic field.

**Typical eruptive sequence and products**

Many vents in the field followed a similar progression of eruptive styles. They commonly commenced with early phreatomagmatic activity as the rising melts encountered shallow groundwater, generating steam explosions and producing large craters (maars) surrounded by tuff cones (tuff rings) of highly fragmented juvenile basalt and accidental tephra. This was followed by or transitioned into typical Strombolian eruptions, somewhat less explosive than the early phreatomagmatic phase, and characterized by intermittent explosive blasts of volatile-rich basaltic lava out of a central conduit, ultimately building the classic cinder cones (scoria cones) that we see dotting the landscape of the Cima volcanic field today. As the supply of volatile-rich basaltic magma waned during each eruption, the central crater of these newly erected cinder cones would start to fill with a lava lake and eventually either overtop the rim, push away a segment of the crater wall, or tunnel through the lower parts of the cinder cone, and the eruption would thus transition into its final strictly effusive phase, feeding extensive lava flows emanating from the vent. Each of these eruptive stages and their characteristic deposits is described more fully below.

**Early Phreatomagmatic Stage**

This stage is well documented at numerous vents (V3, 6-8, 12-14, 16) and likely occurred at most others. In many cases, the associated deposits have either been eroded away or covered by younger scoria and lava. The original craters are inferred to have been up to 1 km in diameter and hundreds of meters deep but are now largely filled by scoria cone deposits of the ensuing Strombolian phase of the eruption. Nevertheless, segments of surrounding low-relief tuff rings are locally well preserved and display the classic stratification, clast compositions, and grain size distributions typical of these types of eruptions. Tuff ring deposits are characterized by a mixture of fragmental juvenile (vesicular basalt) and accidental (country rock) fragments, and are dominated by fine-grained ash and small lapilli-sized tephra, with subordinate blocks and bombs. They are generally very well stratified and typically grade upwards from early erupted light-colored beds dominated by accidental granitic detritus (some that superficially resemble fluvial deposits) to well-stratified dark grey layers dominated by ash and lapilli of highly vesicular juvenile basalt (basaltic pumice). The overall layering, sorting, clast size distributions, arrangement of clasts, and presence of key features (grading, large-amplitude cross stratification, bomb sags) indicate these deposits are a mixture of fallout and surge deposits, occasionally punctuated by the arrival of large ballistic blocks and bombs. The fine grain size (high fragmentation) of the tephra and its wide dispersal suggests eruption columns were many kilometers high (subplinian) during this earliest and most explosive phase of the eruptions, posing a risk to anything within several miles of the vent. We will see nice examples of these kinds of deposits at Stops 2, 3, and 4.

**Strombolian Stage**

Virtually every vent in the Cima volcanic field preserves evidence of Strombolian activity in the form of a cinder cone (or at least remnants thereof). The conical red hills that dot the Cima volcanic field are all classic examples of cinder (scoria) cones—loose aggregates of basaltic scoria lapilli, blocks and bombs that build up around the vent during the early more explosive stage of a basaltic eruption. Their red coloration is a consequence of the oxidation of red-hot basaltic fragments as they flew through the air. These cones have a characteristic architecture with a small central crater encircled by summit rim and smooth outwardly dipping radial slopes of loose scoria. Invariably, the summit rims at Cima are significantly higher on their eastern and southern sides, due to the prevailing northerly or westerly winds causing greater accumulation on the downwind side. Cuts into these cinder cones reveal outward dipping layers lying at angle of repose, with a conspicuous decrease in average clast size with increasing distance from the vent. The cinder cones are up to 800 m in basal diameter and are estimated to have attained heights up to several hundred meters. Strombolian eruptions are driven by explosion of large (magmatic) gas bubbles rising to the surface of the magma conduit, and do not require any interaction with external water. This type of activity ceases once the more volatile rich magma is depleted. For the majority of Cima vents, only a small fraction (<30%) of the total magma volume was erupted in this fashion. The duration of Strombolian activity at each vent is not well known, but was probably weeks to months, though there are examples of modern Strombolian eruptions that have lasted for years.

**Effusive Stage**

The final stage at most Cima vents is the effusion of copious amounts of lava. Initially, this occurs by filling a lava lake in the central crater of the cinder cone. Eventually, the lava finds a way out into the surrounding lowlands by doing one of the following:

- a) Spilling over the rim of the cinder cone and flowing down its flanks (usually on the lower NW side)
- b) Pushing away an entire section of the cinder cone wall (usually on the N or W side)
- c) Propagating radial dikes outward through the cinder cone walls
- d) Tunneling under a portion of the of the cinder cone walls
Of these, (b) seems to be the most common, as evidenced by the abundance of horseshoe shaped cinder cones that open to the W or N, with lava deltas and long runout lava flows emanating from these breaches. It is not unusual to find the missing cinder cone wall more than a kilometer from the vent where it was rafted away by the denser underlying lava flow. Although the surface texture of most of the older lavas has been lost to erosion, it is clear that both pāhoehoe and ‘a’ā lava flows were common, with perhaps a preponderance of ‘a’ā’. The longest runout lengths of individual lava lobes are ~7 km, and average slope angles are consistently ~2°.

Estimated DRE (dense rock equivalent) volumes of magma erupted from individual vents are rather small, ranging from ≤ 1.0 to ~100 x 10^6 m^3 (.001 to 0.1 km^3). The ratio of lava flow to cinder cone DRE volume from individual vents is also highly variable (0.2 to 20), with the large lava fractions associated with the more voluminous eruptions. There is no obvious structural control to the vents, nor is there any systematic spatial migration of vents through time or any kind of time-volume relationship. The Cima volcanic field is fundamentally puzzling. Why would such a geographically restricted and tectonically inactive area be the locus of tiny episodic basaltic eruptions over such a long time? There must be something special about the underlying mantle or the lower lithospheric plumbing of this area!

Stop 1: Inverted topography at 17 Mile point

As you enter the Mojave National Preserve on Kelbaker Road coming from Baker, you first become aware of the CVF at mile 17, where on your left, you pass a conspicuous mesa capped by dark basalt. This basalt is the distal part of a ~200 ka lava flow interpreted to have been erupted from Vent 11, five miles to the east. This stop provides an excellent opportunity to examine an ancient lava flow in cross section and to appreciate how 200 ka of erosion and weathering have modified the local geomorphology. Park in the pullout west of the highway at the south end of the mesa and, using Figure 2 as a guide, take some time to explore the following:

Stop 1A: Directly north of you along the edge of the wash is a cross section of the ~200 ka lava flow, exposing its basal contact. Note, in ascending order:
(a) well-stratified fluvial gravels composed mainly of granitic detritus,
(b) a basal breccia composed of vesicular basalt rubble (much of it coated with tan silt),
(c) a several m-thick massive basalt layer that becomes progressively darker and more vesicular towards the top,
(d) a transition at the top of the cliff from flow-banded vesicular lava to rough, torn up ‘a’ā textured lava. Clearly, most of the original ‘a’ā surface material has been eroded, but you can still be confident that this was originally an ‘a’ā flow by its rubbly base.

Stop 1B: Walk westward to the corner of this escarpment by the highway and climb up on to the lava flow. Here, the edge of the lava is buttressing against a paleohigh of crystalline basement—the same rocks that are exposed in the hills directly across the highway. The top of the lava flow is a relatively flat surface composed mainly of angular pebbles and cobbles of basalt. Though somewhat rough, the surface has very little local relief, and in many places you can see that the surface layer is a mix of massive and vesicular fragments that are arranged in a closely packed mosaic of desert pavement. This surface is strictly aggradational and preserves no vestige of its original ‘a’ā surface (compare with Stop 2A later). If you dig below the surface layer you will find a layer of light-colored silt and sand. The light-colored material...
is mainly windblown (aeolian) decomposed granitic sand and silt and thus could not have been derived from the underlying basalt and in fact is compositionally similar to playa sediment of Soda Lake. Sites like this were influential to our current understanding of how desert pavement forms. In a landmark paper, Wells et al. (1985) suggested that the light-colored underlying soil is composed entirely of windblown silt that lands on the surface and then percolates downward between the pebbles. This process combined with repeated wetting and drying of the clay-rich material gradually builds up a substantial "added" soil horizon, capped by closely packed pebbles derived from the underlying lava. This 'born on the surface' hypothesis was strongly supported by cosmogenic dating of surface clasts, which are the same age as the flow (Wells et al., 1995). Note that this process likely occurred following erosional beveling and removal of several m of 'a`a' surface lava. Now that the lava lies above its surrounding base level, the surface is largely protected from further erosion by the top layer of darkly varnished basalt pebbles, and the main erosional process is the spalling off of blocks around the perimeter of the lava mesa.

**Stop 1C:** Drop back down to the highway and walk north ~500 m to the NW corner of the mesa. The base of the basalt lava is well exposed along the roadcut just below the top of the mesa to your right, and it rests on a thick section of well-bedded fluvial gravels. This relationship, combined with the overall finger-like geometry of the 200 ka lava, indicates that this lobe of lava flowed westward down an ancestral ~500 m-wide alluvial wash whose floor lies now well above the modern washes, and that this lava likely buttressed against the hills of older bedrock across the highway, temporarily damming the main N–S wash. Subsequent erosion has removed the lava dam and lowered the local base level of the modern washes and pediment surfaces by about 20 m, suggesting an average long-term local downcutting rate of ~0.1 mm/yr. At the NW corner of the mesa, you encounter a different basalt lava down at road level. A brief inspection of the edge and top of this topographically lower lava should convince you that (a) its present surface is not nearly as degraded as the mesa-forming one, with substantial local relief and some preservation of 'a`a' texture despite significant cover by younger alluvial sediments, and (b) this lava appears to have flowed around the NW corner of the escarpment formed by the 200 ka lava as it entered the major NW-trending wash. Thus, this topographically lower lava is likely younger than the mesa-forming lava, even though at first glance it might appear to sit "stratigraphically" below it. This is another example of "inverted topography" – wherein younger lavas fill channels at lower elevations and older lava-filled channels now form resistant linear ridges and "finger mesas". The younger lava at this location is not well dated but is likely ≤ 100 ka.

**Stop 2:** Morphology of a young 'a`a' flow and a particularly instructive canyon

*Lat. 35.184371° Long. -115.828114*

V2 and its associated 'a`a' flow is unquestionably the youngest eruption in the entire Cima volcanic field. Numerous attempts to date this lava using different methods (e.g. K-Ar, 40Ar/39Ar, Cosmogenic) have yielded somewhat conflicting results, but the most reliable of these suggest this eruption is ~11–12 ka (Wells et al., 1995; Phillips, 2003). Its youth is indicated by the excellent preservation of the original 'a`a' surface texture, with virtually no soil or vegetation cover, and the general morphology and youthful appearance of the distal flow front concordant with the modern alluvial surface. We will briefly examine a distal part of this lava and then proceed up a canyon that has been recently eroded between the edge of this flow and the V3 edifice to the north (Fig. 3A). The hike up this canyon (Fig. 3B) offers the opportunity to examine different types of eruptive deposits and to contemplate the complex stratigraphic relationships that can be used to reconstruct eruptive histories. The main highlights for each stop along the transect are described below:

**2A:** A quick stop here allows you to see the characteristic front of an 'a`a' flow, with many small lobes and embayments producing a crenulate flow front pattern (Fig. 3A). Climbing up onto the flow makes you appreciate just how rough the surface of 'a`a' is, with all the unstable blocks of rough textured lava making any traverse painstakingly difficult. Note the complete lack of soil development and effective absence of any vegetation on this young lava surface and compare this to what you saw at Stop 1. If you proceed a little southward on top of the lava flow, you will spot some brown to reddish knobs that have a distinctly different surface morphology. Examine these and you will discover an abundance of reddish weathering scoria, including occasional large blocks and bombs up to 50 cm in diameter, some with sculpted aerodynamic shapes, and virtually all displaying the highly vesicular cores characteristic of airborne lava fragments. What are these doing here nearly 2 km from the vent? The simplest explanation is that these are remnants of the western wall of the V2 cinder cone that was rafted away by the underlying 'a`a' flow. This is not an uncommon feature of lava flows emanating from cinder cone vents, and reflects the much lighter average density of scoria compared to the underlying basalt lava.

**2B:** The V3 eruptive center is one of the older centers in this part of the Cima field, based on its highly gullied and degraded cinder cone and the fact that numerous lavas coming from other vents are deflected around it. A scramble up the flank of the V3 cinder cone at this locality reveals a near vertical dike and several irregular sill-like bodies of basalt that clearly cut the scoria deposits. This dike terminates near, and apparently fed, a stack of thin
shelly pāhoehoe lavas that armor the lower southwest flank of the cinder cone, and nicely illustrates one of the ways that lava gets out from a central crater to the surrounding lowlands.

2C: We could only examine the top and front of the V2 ‘a’a flow at stop 2A. Here, recent erosion has cut a deep gully along the edge of the V2 lava, providing a superb cross-sectional view of this young lava on the southern wall of the wash. Notice the crude columnar jointing, the massive (but faintly flow banded) character of the lower 5 m of the lava and how it becomes more vesicular towards the top, and the abrupt transition near the top of the cliff into torn up ‘a’a textured lava crust. This section is fairly thick for a basalt lava, perhaps because here it was filling the deeper portion of an ancestral wash adjacent to the V3 cinder cone.

2D: Exposed in the cuts along the northern edge of the wash are tan well-stratified deposits that dip gently south toward the wash (away from V3). These are remnants of some of the distal tuff ring deposits associated with early phreatomagmatic eruptions from the V3 vent. Some of the layers have conspicuous granitic detritus and angular basaltic clasts, but if you examine them carefully you will see that there is an abundance of finer grained tan fragments of highly vesicular basaltic glass, attesting to a magmatic component. The grain size distribution, distinct stratification, and discontinuous lenses of coarse pebbles all suggest that these are mainly surge deposits. A bit further up the canyon and stratigraphically higher are dark brown to gray, weakly stratified fallout deposits composed almost exclusively of basaltic ash and small lapilli—part of the same tuff ring but with a much greater juvenile component. The original edifice of the V3 tuff ring is no longer evident because it has been largely eroded or covered by debris from the cinder cone, but recent erosion along this wash has exposed its remnants, thereby documenting an early phreatomagmatic eruptive stage for this vent.

2E: Further up the canyon along its southern wall, we have now dropped below the basal contact of the V2 ‘a’a flow and have picked up additional stratigraphic units. Starting directly above the floor of the wash is a 1–2 m section of well-stratified darkish brown tephra that dips to the south and is composed almost entirely of lapilli and ash-sized basaltic pumice with occasional larger blocks (bombs) of basalt—part of the same V3 tuff ring succession we observed at 2D. These are overlain by a thin interval of conglomerate with intervening lenses of coarse sandstone, composed of mixed granitic and basaltic detritus, but clearly sedimentary in origin. These are interpreted to be ancestral alluvial wash deposits—here mainly debris flow deposits—that were deposited on top of distal portions of the V3 tuff ring and immediately prior to the V2 ‘a’a eruption. The very top of this thin sedimentary sequence is reddish (oxidized) just below the contact with the overlying V2 lava. The base of the V2 lava is marked by ~1 m of vesicular rubble—material that cascaded down the
front and was subsequently overridden by the advancing lava.

2F: The plot thickens somewhat a little further up the canyon. A small cliff (dry waterfall) of basalt extends across the wash and clearly overlies and buttresses against tephra deposits of V3 on the northern side of the wash. On the south side, it can be followed eastward a short distance where it apparently terminates but is clearly overlain by the V2 lava. Thus, this basalt cliff represents a new lava that is younger than at least part of the V3 eruption but older than V2. Climb up on top of this cliff and on the south side of the wash you will discover additional fine grained basaltic tephra—both fall out and surge deposits—that rest directly on the eroded surface of the new lava, and appear to dip southward (away from V3). It is puzzling that this lava appears to both overlie and underlie tephra that came from V3. Perhaps the V3 vent was a polygenetic eruption with two stages of explosive activity, and this basalt lava was erupted between the two stages. In any case, the lava is clearly older than the V2 lava and younger than the majority of the V3 edifice.

2G: Finally, a little further up the canyon, we encounter a second dry waterfall/cliff of lava extending across the wash. It is clearly younger than the one at 2F as it sits above that lava, and it apparently flowed westward down virtually the same wash. On the northern wall of the canyon, it rests on and buttresses against the stratigraphically highest tephra deposits related to V3, but on the southern side of the wash it clearly underlies the V2 ‘a’a.

The simplest eruptive history to be gleaned from relations in this canyon is that the oldest recorded eruptions were explosive eruptions associated with the early stages of V3, after which a lava flowed part way down a wash that skirted the southern flank of the V3 edifice. This was followed by additional explosive eruptions, apparently also from V3, and then eruption of a new lava from a source to the east that flowed part way down the same wash and apparently terminated not far beyond the cliff at 2G. A period of renewed erosion and deposition of wash gravels was then followed by eruption of the V2 ‘a’a lava that flowed down the same general path as the previous ones. The final chapter of erosion and downcutting along the boundary between the V2 lava and the V3 vent edifice is the only reason that these complex stratigraphic relationships are exposed for us to puzzle over. It remains entirely unclear which vents the intermediate aged lavas at 2F and 2G came from—presumably vents well to the east of this canyon. Perhaps the most important lesson to be drawn from this hike is that stratigraphic relationships within the Cima field can be rather complicated and not resolvable from Google Earth imagery alone. Broad expanses covered by lava surfaces or alluvial wash sediments are usually not very informative compared to some of the recently incised canyons and washes such as this one.

Stop 3: South flank of V12 – early tuff cone deposits and architecture

A rough 4x4 road (Indian Springs Jeep Trail) bypasses the Stop 2 canyon and continues about 3 miles towards the ENE up the broad wash between vents V3, V11, V12 to the north and vents V4–V7 to the south. At this stop we will examine superb exposures of early phreatomagmatic tuff ring deposits that accumulated on the eastern margin of V12. Park at the bend in the road (35.206067°; -115.785825°) and hike up into the narrow erosional gully that has been carved between the outer flank of the V12 cinder cone and the inner flank of the V12 tuff ring (Fig. 4).

Low on the northern side of the gully you will encounter light-colored, well-stratified beds containing large cobbles and boulders of granite set in a finer grained matrix. Superficially these resemble fluvial/alluvial conglomerate and debris flow deposits, but on close inspection, you will note that the matrix contains abundant angular ash- and lapilli-sized fragments of tan vesicular basaltic pumice (similar to what we saw at Stop 2)—material that is far too delicate to survive significant stream transport. These are the proximal tuff ring deposits associated with

Figure 4. Google Earth oblique view to the northeast of V12 composite cinder cone and peripheral tuff ring, showing the route to and location of Stop 3.
the early phreatomagmatic explosions from V12. Here, the proportion of accidental (granite) fragments exceeds that of juvenile (basalt) fragments, and the beds themselves are mainly surge deposits punctuated by occasional large ballistic bombs. Many of the larger granite boulders clearly deflect the underlying layering producing classic bomb sags, and many are rounded, suggesting that the vent may have intersected pre-existing stream channel deposits, or perhaps they were rounded by repeated ejections and abrasion in the vent area. The stratification here dips gently northward toward vent 12, but looking eastward up the gully, it is clear that higher beds dip outward (southward) in what, from here, appears to be a major angular unconformity.

Continue up the gully, and several things will become apparent. The change in dip direction occurs gradually rather than abruptly across an unconformity, as individual beds can be traced from north-dipping to south-dipping across a conspicuous antiform. This transition from inward dipping beds on the inside of a tuff ring to outward dipping beds on the outside is typical of all tuff rings, and reflects the fact that during tuff ring growth, successive deposits broadly mantle the underlying slope. Because there is always an optimal distance from the central vent where net accumulation rate is maximized, over time the crest of the tuff ring will build up more rapidly at this distance, and the inner and outer slopes will steepen, with mantling deposits that mimic the underlying slopes.

Climb stratigraphically up through the tuff ring deposits on the south wall of the gully to the crest of the tuff ring and you will note a gradual transition from light-colored granite-rich horizons to darker gray layers composed of an increasing proportion of basaltic ash and lapilli, with only sporadic granite clasts. These thinly layered strata include both fallout and surge deposits and their high degree of fragmentation suggests that they record continued phreatomagmatic activity, albeit with a central vent that now had largely been cleared of country rock. Some of the highest material on the tuff ring might also represent distal fine-grained tephra from the earliest stages of the subsequent Strombolian eruptions. From the crest of the tuff ring, look northward to see a columnar-jointed cliff of flat-lying basalt lava perched well up on the flank of the red cinder cone. What is flat-lying basalt lava doing way up there? This is a remnant of a lava lake that once filled the moat between the tuff ring and cinder cone, analogous to a situation that we will explore in greater detail at Stop 5.

Stop 4: Eruptive history of V5, V6, and V7 vents

Turn off Kelbaker Road, and head northeast along the Aiken Mine road for 2.3 miles. After passing a broad composite lava field surrounding the highly dissected cinder cone of V4 on your left you have arrived next to the cluster of V5, V6, and V7 vents (Fig. 1). This is a complicated area, with coalescing lava flows that emanated from different vents and with ambiguous stratigraphic relationships. Here we focus on the local sequence of eruptive events related to these three vents, all of which are clearly younger than V4, as their lavas flow around that edifice. From the road, hike northwestern to the south rim of V6 (35.185280°; -115.775129°). Along the way, you first cross a relatively youthful looking lava adjacent to the wash, with locally well-preserved pāhoehoe surface. Next, climb up to the crest of a low ridge with conspicuous bands of darker and lighter material. Surprisingly, the light bands are composed almost entirely of granitic detritus with boulders up to 1 m in diameter. A quick inspection reveals that they are in fact locally derived, weathering out of stratified outcrops. How did this granitic material get way up here on the flanks of the V6 cinder cone? The answer is that these are the coarse tephra deposits from violent phreatic and phreatomagmatic explosions, and you are standing on the remnants of what appears to be a large composite tuff ring. This low arcuate ridge continues hundreds of

Figure 5. Google Earth oblique northward view along the V5, V6, and V7 cinder cones showing route to Stop 4, position of composite V7 tuff ring, and flow trajectories of major lava lobes sourced from V7. See text for discussion.
meters to the NE where it appears to be broadly concentric around the V7 cinder cone, while the segment that trends off toward the SW appears to mantle the eroded SE flank of V6 and NE flank of V5 (Fig. 5). I interpret this tuff ring to be related to the early phreatomagmatic stage(s) of the V7 eruption, with some of the ejecta here mantling the older V5 and V6 cinder cones. Note that this arcuate ridge was interpreted by Wilshire (1992) to be a separate vent (V6B) —an interpretation with which I do not agree. The double crescent shape and large diameter indicated by this tuff ring remnant suggests that there may have been multiple (nested) maars and a particularly vigorous phreatomagmatic stage to the V7 eruption.

The view from the crest of the V6 cinder cone reveals additional details. It appears that Strombolian activity proceeded from V5 to V6 to V7, as the resultant cinder cones are successively less degraded, and lavas emanating from the younger ones appear to be deflected around the older edifices. It is also apparent that each successive eruption was more voluminous, given the progressively larger size of their respective cinder cone and lava field. Finally, this view sheds light on how lava exited the V7 vent. The inside of the northern part of the tuff ring (northeast of this point) contains lava that forms a small cliff coincident with the rim of the tuff ring and a triangular wedge that slopes back towards a point on the lower SE slope of the V7 cinder cone (Fig. 5). I infer that lava exited from a tunnel at about this location (no longer exposed) by pushing its way through the lower portion of the V7 cinder cone during fairly early stages of the eruption and filling a lava lake inside the tuff ring. Eventually this lake spilled over the rim of the tuff ring, carving a bit of a channel in the process (still visible today) and feeding a long skinny lobe of pāhoehoe lava that flowed at least 3.5 miles, first towards the SW then to the W, following the wash that skirted around the edge of the earlier V4, V5, and V6 vents and their associated lava fields (Fig. 5). This is the lava that we encountered at the beginning of our hike. An even greater amount of V7 lava subsequently flowed out through a breach in the western upper wall of the cinder cone, forming a broad lava delta on that flank and extensive lava fields in the lowlands to the west. It is also evident from here that the three vents (V5, 6, 7) are very well aligned in a NNE direction.

I speculate that V5, V6, and V7 are closely related and that perhaps they erupted in rapid succession, progressing towards the northeast along the same underlying fissure, with V7 encountering a large reservoir of shallow groundwater and also being by far the largest of the 3 eruptions.

Stop 5: Vent 13—Outer tuff ring, inner scoria cones, and draining of an immense lava lake

Caravans of weekend warriors in decked-out overlanding rigs make their way to a well-marked but not particularly noteworthy lava tube situated on the western flank of Vent 13, making this perhaps the most visited spot in the entire Cima volcanic field. Despite the crowds, this area is well worth a visit, as it tells an interesting story above and beyond the presence of a lava tube.

Park at the designated spot for the lava tube (Fig. 6), hike up to the lava tube entrance and continue eastward up to the top of the ridge (Fig. 6, Stop 5A). You can stop and inspect the lava tube on your return. You are now standing on the crest of a tuff ring edifice that was created during the early phreatomagmatic stage of the V13 eruption. This is perhaps the largest and best-defined tuff ring in the entire field, with good exposures of the deposits on its eastern flank in the wash adjacent to the Aiken Mine road. Vent 13 is one of the youngest eruptive centers at Cima, with radiometric ages on associated lava ranging from 27 to 140 ka but mostly clustered around 30 ka (Wilshire, 1992). Looking to the E and NE from here, you can barely make out portions of the far edge of this broad ( > 1 km wide) tuff ring, as your view is mostly obscured by the reddish hills of the cinder cone complex that fills in the central part of the original crater. Note the deep peripheral moat between the inner slope of the tuff ring and the outer slopes of the internal cinder cone and imagine how much deeper this maar might must...
been prior to filling with the Strombolian stage scoria cone! A couple of puzzling observations emerge. First, the surface you are walking on does not look anything like typical tuff cone deposits, but rather is composed of large blocks of vesicular basalt lava. If you look around, you can see definite outcrops of subhorizontal shelly pāhoehoe lava remnants just below the inside rim of the tuff ring and across the moat at similar elevations on the flanks of the cinder cone. What is this lava doing way up here? The simplest explanation is that following Strombolian activity, lava effusions from the vent filled the entire moat area with a large lava lake, until it started spilling over the topographically lowest western edge of the tuff ring and spreading out as broad shelly pāhoehoe lavas on the plain below. This explains why the western portion of the tuff ring appears to be so much darker on Google Earth imagery than the rest of the tuff ring—it is blanketed by heavily varnished black basaltic talus that represents this overspill lava.

A second question emerges; if this entire depression within the tuff ring was once filled with lava, where did it all go? Look back at the position of the lava tube. The lava tube starts at the stairway entrance and extends downhill about 100 m directly away from the tuff cone, terminating at both ends due to collapse of the roof. Its position and orientation suggest that it originally continued beneath the western part of the tuff ring, and it is easy to imagine that this tube was created by tunneling of overpressured lava within the lava lake beneath the tuff ring and served as the drain pipe for the lake and subsequent effusions from Vent 13.

On your return down the hill you can take a minute to explore the lava tube (5B) with its thin roof of shelly pāhoehoe lavas, several skylights, a few lava stalactites, but mostly a lot of dust and accumulations of blocks from partial collapse of the roof. Where do you think all the light-colored silt on the floor of the tube came from? Back on the surface, continue walking westward past the jeep trail to where a subdued valley-like depression with raised edges opens toward the SW, directly on line with but about 70 m west of the last skylight into the lava tube (5C). This is a classic lava channel, and is interpreted to mark where the lava tube exited to the surface. The lava channel widens downslope to ~100 m, has well-preserved marginal levees, and can be traced for ~1.5 km to the SW, where it fed distal lavas from this prolific vent. Most of the lava from this vent was pāhoehoe and some clearly flowed more than 7 km down the broad wash to the SW. As you walk around this area, you should be able to find excellent examples of pāhoehoe surfaces, tumuli, pressure ridges, etc. attesting to its youth. Based on its preserved deposits, V13 had a vigorous phreatomagmatic stage, a relatively minor Strombolian stage, and then produced copious amounts of highly fluid pāhoehoe lavas (>90% of total erupted volume).

Stop 6: Vent 14—Aiken Cinder Mine, mantle xenoliths, lava dams, and geomorphic evolution

Lat. 35.229159°  Long. -115.726125°

There are a number of excellent reasons to visit the Aiken Cinder Mine but be forewarned, the access road is very rough and requires high clearance 4x4. Cinder (scoria) was mined here for road grade and other construction purposes, by excavating part of the northern flank of the V14 cone and sieving the tephra into various size fractions. Mining operations ceased around 1990, but much of the old mining equipment is still present, and the various mining cuts provide superb exposures of the internal architecture of a cinder cone. This vent is notable for the abundance of bombs cored by crustal and mantle xenoliths, and for how the growth of the cinder cone and subsequent effusion of 'a'a lava fundamentally modified the local drainage system.

Begin your visit by walking around some of the mine workings on the northern side of the cinder cone (Fig. 7). The various cuts provide exceptional views of the characteristic layering and grain size distribution of fallout deposits associated with Strombolian eruptions. Lapilli-sized (2–64 mm) fragments of scoria with subordinate larger blocks and bombs are neatly stacked in layers defined by subtle variations in clast size. The scoria represents originally incandescent fragments of lava,

Figure 7: Annotated Google earth image illustrating geologic relationships in the vicinity of the Aiken Cinder Mine (V14).
that were hurled in the air hundreds of meters by the rapid ascent and bursting of large gas bubbles in the vent, rapidly cooled and solidified while airborne, and then fell back to earth and accumulated around the vent. Innumerable blasts like this gradually built up a conical edifice over a period of weeks to months, with a central crater and outward dipping layers of ejecta lying at the angle of repose. Examine some of the larger bombs of scoria that are broken in half and you will see a characteristic zoning from an oxidized and somewhat molded looking outer surface that resembles breccia inward through an outer rind with sparse tiny vesicles, to an inner core with abundant large vesicles. This type of zonation is definitive proof that as the bomb sailed through the air, its outer portion rapidly quenched and solidified while its interior remained molten and continued to vesiculate. Some bombs are up to 1 m in diameter with beautiful aerodynamically sculpted shapes. If you pick up some of the grapefruit sized bombs, most will feel very light, but a few will feel exceptionally dense. Break these open and you will find that the light ones are highly vesicular scoria, the medium density ones are non-vesicular basalt lava, and the rare very dense ones are bombs that are cored by ultramafic and mafic xenoliths coated with a thin veneer of basalt. These xenoliths were brought from upper mantle and lower crustal depths by the rapidly ascending basalt magma and provide important clues about the composition and character of the lower crust and upper mantle beneath this region (e.g. Wilshire, 1991). The mantle xenoliths are mainly pyroxenite and websterite and crustal ones range from granite to gabbro. Given the large size of some of the ultramafic xenoliths, and much higher density of pyroxenite relative to basalt magma, one can demonstrate using Stokes law that these basalt magmas likely ascended from their mantle source region (~ 50 km depth) to the surface in under 12 hours.

A climb to the top of the cinder cone on its NE flank will reward you with panoramic views of the surrounding area and provides an excellent vantage point to reconstruct the sequence of local eruptive and geomorphic events. Several kilometers to the east of your position is a conical peak (Winters) formed by the highly degraded cinder cone of Vent 15. Emanating from this vent and sloping toward you (away from the crest of Cima Dome) are several obvious lava lobes with mature erosional surfaces that now form elevated finger mesas. These lavas are ~700 ka and can be traced up to 5.5 km eastward where they reached the major wash NW of V13. At first glance, it is tempting to assume that these finger-like mesas of lava once formed a single large continuous sheet. That is probably not the case, because towards the headwaters of the canyons between these lava lobes are bedrock hills that clearly stand higher than the adjacent lavas, indicating that they were in separate channels. This, combined with the fact that the lavas themselves consistently rest on sections of alluvial wash gravels, suggests that these were distinct finger-like lobes of lavas flowing down separate washes at the time. Now there are deep canyons between these finger mesas, as subsequent erosion has preferentially lowered what was previous high ground between the different lobes of lava, thereby inverting the topography.

From this vantage point it is also evident what transpired at V14. This vent is clearly younger than the V15 eruptions, given the more youthful expression of the V14 cinder cone and the well-preserved 'ā'a lava surfaces. Furthermore, 'ā'a lava emanating from V14 rests directly on the eroded surfaces of distal V15 lavas just S of the cinder cone (Fig. 7). It is also apparent from here that the effusive stage of the V14 eruption involved breaching and rafting away of the SW wall of the cinder cone leaving a U-shaped amphitheater, with some of the effluent 'ā'a lava flowing westward in the prevailing downslope direction, but some also flowing eastward and then northward, wrapping around the cinder cone and filling the lower portions of the canyons between the V15 lava lobes. In effect, the V14 cinder cone and its lava has dammed up these canyons, leaving the peculiar spectacle of deeply incised major canyons upstream ending abruptly into playas like flats filled with sediment just a few km downstream.

**Stop 7: Indian Springs—the great unconformity**

*Lat. 35.233224°  Long. -115.814495°*

At previous stops, we have examined lava flows deposited on older Quaternary alluvial sediments. This stop provides an opportunity to examine a xenolith-rich lava that flowed down a bedrock channel cut into Proterozoic gneiss. Turn east off Kelbaker Road onto an unmarked dirt track (35.215080°, -115.876571°) and proceed northeastward about 3.0 miles to where the track enters a broad sandy wash. You might want to walk from here unless you have...
4x4, or you can drive a bit further. Either way, you want to continue up Indian Springs wash an additional ~0.92 miles to where a narrow side canyon enters from the north (Fig. 8). En route, you will pass excellent outcrops of Proterozoic banded gneiss (inferred to be ~ 1.7 Ga), cut by intricately folded leucogranite veins and dotted by many barrel cacti—all forming an aesthetically pleasing desert landscape.

A short hike up into the side canyon provides you with beautiful cross-sectional views of a Quaternary basalt lava lobe that was erupted from a composite vent 3 km to the east, here resting directly on the banded gneiss. Key takeaways include:
- This lava is filling a ENE–WSW oriented bedrock channel, with clear buttressing relations against paleocanyon walls of gneiss to the north and south.
- The vesicular rubble at the base of the lava demonstrates this was an ‘a’a flow, though it now has a smooth desert pavemented surface.
- If you examine the large blocks in the floor of the modern canyon, you will find abundant crustal (gabbro and diorite) and lessor mantle (pyroxenite) xenoliths as well as occasional large plagioclase xenocrysts—a feature of many of the eruptions in the Cima field (e.g. Stop 6).

Take a moment to consider the vast amount of geologic time that is missing across this unconformity (nearly half of Earth’s history) and imagine how much a future geologist would be missing (effectively all of Cordilleran history) if this was all they had to work with.

References


Young basalt fields of the Mojave Desert

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ABSTRACT—Basalt, a mafic volcanic rock common in mid-ocean islands and in several continental settings, is melted from upper mantle rocks in many cases and thus provides information on mantle conditions. Basalt lava fields, some decorated with scoria “cinder” cones, are scattered around the Mojave Desert. Only a few basalt fields have been well studied, so we undertook a compilation of basalt fields that are younger than ~12 Ma to examine space-time patterns. The Cima volcanic field is unique in having eruptions that span ~7.5 m.y., including the youngest eruption in the Mojave Desert at ~12 ka. Other fields probably erupted over short timespans of decades to hundreds of years based on analogy with modern eruptions, with a few exceptions of fields with perhaps 1 m.y. timespan. We find that all basalt fields except Cima are restricted to the active eastern California shear zone, and many lie directly on active faults, indicating a direct relation between faults and volcanism. Area and volume of lava is greatest for those fields associated with dextral faults, which may be attributed to less compressive stress across those faults and less resistance to buoyant rise of magma, as compared to sinistral faults. Xenolith-bearing basalts that include both mantle and deep crustal rocks are known in the eastern San Bernardino Mountains area, Dish Hill and Deadman Lake fields near Amboy, and Cima volcanic field.

Introduction
Young basaltic lavas and cinder cones have been noted in the Mojave Desert by many geologists and travelers. Basalt lavas are mafic magma that requires a substantial mantle component as a source for melting, and a path for ascent to the surface. As a result, basalt can inform on past mantle and crustal conditions. Geochemical and isotopic studies of basalt are powerful for inferring mantle source characteristics but are not treated in this study; readers can refer to Katz and Boettcher (1980), Hughes (1986), Glazner et al. (1991), Waits (1995), Farmer et al. (1995), and Luffi et al. (2009). As the geologic study of the Mojave Desert has progressed, much has been learned about basalts, but that information has not been compiled, including three fields that we have studied recently. We undertake a compilation of locations and ages for basalt fields and provide preliminary results in this study.

We compiled a map of young (<12 Ma) basalt fields in the Mojave Desert, which we define as framed by the San Andreas and Garlock faults and limited by the Colorado River on the east. The southern limit is at the southern border of Joshua Tree National Park, south of which the basalt ages are early Miocene or older (Carter et al., 1987). Several well-known fields (Fig. 1) are comparatively well studied, including the Cima volcanic field, and the basalt fields at Pioneertown, Pisgah, and Amboy are known to travelers because they lie along major transportation routes. Other basalts that have been studied include northern fields: Dish Hill, near Amboy; Black Mountain, northeast of Harper Lake; and Bicycle Lake within the Fort Irwin National Training Center; and include southern fields: Fry and Ruby mountain fields north of Yucca Valley. Less well studied are scattered cinder cones and small flow remnants of the Old Woman Springs field, several within Joshua Tree National Park, those in the San Bernardino Mountains, and basalt near Deadman Lake within the Twentynine Palms Marine base, as well as a few that are in remote locations such as China Lake Naval Weapons Station (CLN).

A recently published review of 37 Quaternary basaltic fields of the American Southwest, which did not include the basalt fields of the Mojave, summarizes styles of eruptions (effusive, explosive magmatic, and phreatomagmatic), resulting deposit and landform morphology, inclusion of xenoliths, geochemistry, geochronology, regional and local distributions of fields and vents within fields, and the evolution and volcanic hazards of these fields (Valentine et al., 2021). The paper describes fields where many of these topics have been studied, but points out that many fields are minimally studied (at best). Valentine et al. (2021) focused on Quaternary volcanoes and volcanic fields, whereas our study of young basalt fields in the Mojave Desert includes deeper time and has a more restricted geographic region and tectonic association, but both address the importance of detailed physical volcanology, geochemistry, and geochronology in understanding the volcanic history and potential research opportunities.

Results
The basalt fields shown in Figure 1 are a subset of those identified by T.W. Dibblee, Jr., as Quaternary basalt (numerous maps at National Geologic Map Database (usgs.gov)). Dibblee’s criteria included little degradation of the lava flows as well as flat-lying orientation. The latter criteria fail in the cases of older basalts that have not been
Figure 1. Shaded relief map of the Mojave Desert showing basalt fields compiled from numerous map sources (references available on request). Colored polygons show the total range of basalt occurrences and inferred buried or eroded extent of each field; in some cases actual outcrop is too small to be discernible. Not all fields have been well defined and some outlines may change with more study. Locations north and south of Eagle Crags have sparse data, and are inferred from aerial photographs. Triangles show basalt identified in boreholes. Reported ages that have sample locations shown as dots, with age in Ma. Heavy gray lines are young faults.
deformed, as many subsequent dating studies have shown (e.g., Miller et al., 2014). All but a few of Dibblee’s mapped Quaternary basalts have been confirmed or denied with age determinations, and with the exception of basalt within CLN, we feel confident that the basalt fields shown in Figure 1 capture all those Dibblee mapped that are younger than 12 Ma, as well as several outside of the area he mapped. The fields form a northwesterly oriented swath in the central part of the area we investigated. Most fields lie within 50 km of adjacent fields, but there are wider gaps near Twentynine Palms and Barstow. Many of the basalt fields include the lava field and one or more scoria (“cinder”) cones, but some fields do not have identifiable cones, and this might result from that part of the field being eroded more efficiently than the lava field (e.g., Valentine, 2022). Where scoria cones are preserved, they consist of scoria and spatter deposits resulting from explosive eruptions. Basalt fields consist mostly of lava flows that spread laterally, generally being controlled by topography but in some cases ponding deeply enough to break out across former high ground (e.g., Broadwell Mesa, Black Mountain fields). The field surfaces typically are higher elevation near explosive vent areas, where lava spatter and agglutinate deposits, as well as successive lava flows, build topography quickly, but can also have had effusive eruptions along linear vents with little topographic relief in the source areas.

The basalt fields in many cases have at least one radiometric or cosmogenic date, and most appear to be younger than ~8 Ma based on rates of weathering and erosion of scoria cones and flows at Cima (Dohrenwend et al., 1984) where dated flows of 7.5 Ma and younger exist (Turrin et al., 1985; Wilshire et al., 2002a, b, c, d). Cima ages fall into three periods: an early 7.5–7.0 Ma set, a middle 6.2–3.2 Ma set, and a late <1.0 Ma set. These groups are chemically distinct, as well (Farmer et al., 1995). The youngest flow has been of interest and has been dated by many methods. The most recent work has been by cosmogenic methods, with 3He yielding $13 \pm 3$ ka (Wells et al., 1995) and 39Ar age of $12 \pm 2$ ka (Phillips, 2003). Spatially, the early set of flows is restricted to the southeastern part of the field, whereas the middle set is spread over the entire field, and the late set is restricted to the south half of the field. Dated dikes in the gap between flows of the northern and southern parts of the Cima field suggest that flows may have once existed in that area (Wilshire et al., 2002b). Cima appears to the only field with a wide range of ages, and xenoliths are present in lava erupted throughout the age span.

Most fields other than the Cima field are monogenetic, each with a single set of morphologically similar flows and a single vent area, all of which probably formed over a very short period of time, such as one century. Pioneertown field is an exception, with eight thick olivine basalt flows (Neville and Chambers, 1982) and one interbedded tongue of the Old Woman Sandstone (Sadler, 1982a) that indicates encroachment of fluvial deposits on the constructional landforms of basalt flows, requiring longer time for the field to develop. In addition, phenocryst composition varies among the flows. Pioneertown field has two possible vent areas, each with positive topography and red oxidized material determined from remotely sensed images; these may represent scoria cone remains, both of which differ from the remainder of the field with its planar black surfaces. One possible vent area (Chaparrrosa Peak) is southwest of the main area, and one is in the northeast end of the northern mesa. West of Pioneertown the Cienega Seca field (in which we combined three small basalt exposures in the Santa Ana River drainage area) is represented by as many as six basalt flows interbedded with sandstone (Sadler, 1982a), and may represent eruption over several hundred thousand years. Alternatively, the flows may belong to separate fields that have been largely destroyed by subsequent faulting. The Pisgah field (−23 ka; Phillips, 2003) also appears to be an exception with three distinct eruption phases for which scoria cones and lava flows are distinguished on the basis of phenocryst composition, and each phase had different locations of pyroclastic and effusive vents (Wise, 1966). Wise (1966, 1969) did not speculate on the amount of time to petrologically develop these distinct sequences. Acknowledging the stratigraphic and petrologic complexity of the Pisgah field, based on paleomagnetic data (D. Champion, personal communication 1990 in Glazner et al., 1991) the eruptive cycle was completed in less than about 10–20 years (Glazner et al., 1991), thus making this short time of eruption essentially a monogenetic system. The Old Woman Springs field has as many as 14 scoria cones (Sadler, 1982b), but lava has not been divided into separate flow units; it may represent a monogenetic system with multiple vents or explosive edifices above springs. The Deadman Lake field includes several cones and flows that are chemically similar to the Dish Hill lavas (e.g., Wilshire et al., 1980; Howard, 2022), but lack of dating precludes evaluation of the longevity of the field.

Outside of the Cima field, dated fields range in age from ~12 Ma to 23 ka as summarized in Table 1. From oldest to youngest, these dated fields are described briefly below. Areas with the oldest fields are near Eagle Crags and in Joshua Tree National Park (fig. 1). Two basalts form mesas to the west and north of the Eagle Crags volcanic field. The well-developed, gently sloping mesas in the northwest (Black Hills field) are ~7.5 km long, up to 4.5 km wide, and capped by basalt with an 40Ar/39Ar age of 11.66 ± 0.12 Ma (Andrew et al., 2014). The basalt overlies basaltic andesite with an 40Ar/39Ar age of 19.63 ± 0.64 Ma (Andrew et al., 2014), which in turn overlies either tuff or Mesozoic plutonic rocks. The lower sequence is correlated with the Eagle Crags volcanic field to the southeast. The high point of the mesa appears to be the vent area for the basalt (Miller et al., 2022). A dark, narrow ridge trends nearly due south from the inferred vent across underlying rocks and may be remnants of a feeder dike. Another
basalt north of Eagle Crags was mapped by Jennings et al. (1962) as Quaternary or Tertiary. It is more eroded than the former basalt and could be older than the 12 Ma timeframe for this paper, but it is undated to our knowledge. Both of these basalts lie within CLN and are little studied. In the Lava Mountains to the northwest of these basalts, more basalt that was grouped with the Black Hills basalt by Andrew et al. (2014) has ages that are older, in the 12–13 Ma timeframe, and as a result are not included in the our compilation. Evaluating whether these separate fields are part of a larger and long-lived field requires further study. In Joshua Tree National Park, two dates for the Pinkham Canyon field suggest that it and the nearby Washington Wash field may be 12–15 Ma (Carter et al., 1987).

A K-Ar date on the Fry Mountain field indicates a ~8.9 Ma eruptive age for it, and, by extension to nearby compositionally similar xenolith-bearing basalt with similar xenolith compositions, the Ruby Mountain field (Neville et al., 1985). The basalt of the Antelope Creek field also contains ultramafic nodules and may also be grouped with the ~ 9 Ma fields at Fry Mountain and Ruby Mountain. These fields lie north and west of the Pioneertown field. Several K-Ar dates suggest that the Pioneertown basalt is about 7–9 Ma (R. J. Fleck, written comm., 2021). West of the Pioneertown field, similar 7–9 Ma ages for basalt samples belonging to small remnants of flows in the Cienega Seca field (Miller et al., 2014) may point to multiple eruptions over a short time interval, outlining a broad pediment (Yule and Spotila, 2010). Remnants of several small fields in eastern Joshua Tree National Park area are about 6–7 Ma, including those at Pinto Well and Eagle Mountain (Langenheim and Powell, 2009; R.J. Fleck, written comm., 2021).

Farther north, three basalt fields with ages between 5.4 and 3.9 Ma are comparatively well dated: Broadwell Mesa at 5.4 Ma (Buesch and Phelps, 2016; Phelps et al., 2017); Bicycle Lake at 4.5 Ma (Buesch et al., 2022); and Black Mountain at ~3.9 Ma (Oskin and Iriondo, 2004; P.B. Gans, written comm., 2021). The ages of Broadwell Mesa and Bicycle Lake basalts stem from recent study at U.S. Geological Survey (USGS) labs to supplant previous, somewhat older, ages with no accompanying data (Schermer et al., 1996; Brady, 1992). Two basalt fields with only single dates are similar in age to the group of three described above. Several small fields, or remnants of larger fields, are scattered near Old Woman Springs; their only date is 4.9 Ma (Miller et al., 2014). The Ash Hill field east of Ludlow is flat-lying with lava rimming two adjacent eroded pyroclastic edifices in the north (probably phreatomagmatic) and flows outlining a linear effusive vent area in the southwest. The flows are significantly eroded at the south margin, an observation that accords with the K-Ar age of 5.3 Ma (Miller et al., 2014).

Six fields other than Cima have been dated as Pleistocene, from youngest to oldest: 1) Pisgah, 23 ka, and 2) Amboy, 79 ka (Phillips, 2003); 3) Lead Mountain, ~360 ka (Howard, 2002); 4) Lavic, ~750 ka (Oskin et al., 2008); 5) Pipkin, ~770 ka (Oskin et al., 2007); and 6) Dish Hill, ~2.1 Ma (Wilshire and Trask, 1971 and Wilshire et al., 1980). The flows of Amboy and Pisgah are only slightly buried at locations where lava flows overran playas. The Sunshine field near Pisgah appears to be about the same age based on erosional characteristics. The Lead Mountain field is largely undissected (Howard, 2002). In contrast, the toes of the Pipkin and Lavic flows are buried by alluvium. Flows of Dish Hill and nearby undated flows are buried progressively southward toward the axis of a wide alluviated valley. The undated Deadman Lake cones and flows are similar in chemistry to the Dish Hill flows and possibly the same age. Howard (2022) proposes that the Deadman Lake and Dish Hill fields are buried where they connect in the intervening valley, citing evidence in the form of short-wavelength magnetic anomalies in the valley as indicating possible cones or feeder vents. Bedford et al. (2012) mapped two dark colored hills in the intervening valley as basalt.

Age patterns for the young basalt fields are difficult to interpret in detail due to the uncertain quality of much of the age data. With the exception of the Cima field and the

<table>
<thead>
<tr>
<th>Age interval</th>
<th>Basalt field</th>
<th>Number of dated rocks</th>
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<tbody>
<tr>
<td>12–8 Ma</td>
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<td></td>
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<td>1</td>
</tr>
<tr>
<td>7.9–6 Ma</td>
<td>Pinto Well</td>
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<td></td>
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<td>Cienega Seca</td>
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<td>5.9–4 Ma</td>
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<td>2.59–1.00 Ma</td>
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<tr>
<td></td>
<td>Pisgah</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>Sunshine</td>
<td>0</td>
</tr>
</tbody>
</table>

Notes: Ma, million years; ka, thousand years
Cima field is not listed; it ranges from ~7.5 Ma to 12 ka.
Black Hills field in the north, the oldest fields generally are in the south. Fields between 6 and 3 Ma span much of the northern third of the distribution along with one field at Old Woman Springs farther south. The Quaternary fields are in the center of the distribution, spread between Pipkin on the west and Amboy on the east.

With regard to quality of age control, dates on the Bicycle Lake basalt are instructive. Schermer et al. (1996) reported $^{40}\text{Ar}/^{39}\text{Ar}$ ages for two places in the field, 5.50 ± 0.20 Ma and 5.57 ± 0.26 Ma. Data for these heating experiments were not published, making it difficult to evaluate them. Miller and Yount (2002) published K-Ar ages at a different location in the field, with an average age for two runs on the same whole-rock separate of 3.3 ± 0.2 Ma. Another unpublished K-Ar date on the field was 2.9 ± 0.2 Ma (J.K. Nakata, written comm., 1995). The presence of stratigraphically overlying 3.4 Ma ash beds at the latter site (Miller and Yount, 2002) indicated that the K-Ar ages might underestimate the age of the basalt. A new $^{40}\text{Ar}/^{39}\text{Ar}$ date yielded a 4.55 ± 0.07 Ma age (Buesch et al., 2022). This latter age uses more recent protocols for material size and adjusts for alpha recoil, and therefore it is the most reliable date for the field. K-Ar and multiple $^{40}\text{Ar}/^{39}\text{Ar}$ dates also exist for the Black Mountain basalt. Three $^{40}\text{Ar}/^{39}\text{Ar}$ isochron ages by Oskin and Iriondo (2004) are 3.56 ± 0.08, 3.74 ± 0.05, and 3.77 ± 0.11 Ma, with a weighted mean age of 3.7 Ma. They are somewhat younger than the 3.93 ± 0.02 and 4.1 ± 0.3 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ ages obtained by P.B. Gans (written comm., 2021), for which a weighted mean age is 3.93 ± 0.15 Ma. Both sets of $^{40}\text{Ar}/^{39}\text{Ar}$ ages disagree with the previous 2.55 Ma K-Ar age of Burke et al. (1982). The reliable parts of the plateau spectra for the younger two of the Oskin and Iriondo (2004) samples can be interpreted as ages closer to 3.75 and 3.78 Ma. The 3.75 to 3.93 Ma age range of the best dated samples indicates eruption over more than 100 ka, which is in conflict with the observation that no eolian material, soil, or weathering has been observed between flows. On this basis we consider the weighted mean age of 3.93 ± 0.15 Ma as the best age for the basalt field at this time. Reconciliation of monitor mineral ages and other factors such as materials analyzed are needed to further evaluate the $^{40}\text{Ar}/^{39}\text{Ar}$ ages.

Acknowledging that basalt dates have inherent uncertainties that are difficult to evaluate, and that some basalt fields are undated, we find one pattern in ages that we believe is supported. Ages are clustered as four eruptive pulses spanning from ~9 to 6 Ma, from 5.4 to 4.7 Ma, from 3.9 to 3.1 Ma, and younger than ~2.1 Ma (Table 1). These pulses are significantly driven by the shear number of K-Ar dates for the Cima field, but the added dates from other fields help to refine the ages for eruptive pulses beyond the three pulses seen at the Cima field alone. The youngest group includes a 2.03 ± 0.12 Ma K-Ar age on amphibole and 2.1 ± 0.2 Ma fission track age for apatite in a granitic inclusion within Dish Hill (Wilshire and Trask, 1971; Wilshire et al., 1980); others in the youngest group that are dated are younger than 0.77 Ma as summarized above.

Spatially, the basalt fields define interesting patterns. The two most obvious patterns are that the individual fields are widely scattered, which can be stated as a lack of clustering that might be the result of a large magmatic system, and lack of nearby fields with similar ages that might represent the footprints of a region of the mantle that produced the basalts. All but the Cima field lie within the active eastern California shear zone (ECSZ) as defined by Dokka and Travis (1990). The Cima field lies close to the east margin of the ECSZ where parallel inactive strike-slip faults have been mapped (Skirvin and Wells, 1990; Miller, 2012). The western and eastern Mojave Desert adjacent to the ECSZ lack young basalt fields except close to the Las Vegas region where basalts that are more properly grouped with the central Basin and Range province are not shown on Figure 1. Many basalt fields lie on one or more active ECSZ faults and in many cases the scoria cones coincide with faults. The Pisgah, Lavic, Sunshine, and Broadwell Mesa fields are compelling examples. These observations may be explained by faults providing cracks through the brittle crust and thereby forming preferential intrusion and orientation loci. Additionally, deep shear zones associated with the faults may promote generation of basalt (Valentine et al., 2021). The large fields are restricted to faults with northwest strike (dextral) and several of the large fields lie along a broader WNW trend from Amboy to Pisgah. In the Pisgah field, details of eruptions display yet more linear trends. Vents for each of the three eruption cycles form a NNW line, with successive vents progressing to the SSE (Wise, 1966). Each cycle began with formation of scoria cones (with minor lava flows) in the NNW. Each scoria cone eruption was followed by effusive vents in the SSE where fissure vents or clusters of vents were aligned to the SE or SW, probably along feeder dikes. The WNW trend from Amboy to Pisgah fields is associated with topographic lows, possibly caused by right steps in the dextral faults (releasing bends) that would further enhance intrusion. Furthermore, small fields along east-striking sinistral faults may owe their diminished size to increased compressive stress across those faults, limiting intrusion by dikes. An estimate of eruptive volume for fields associated with dextral vs. sinistral faults indicates nearly five times greater average volume for the basalt fields associated with dextral faults, and that is reflected in the map pattern of Figure 1. Interestingly, arguments can be made that basalt magma, which has a density of 2.55 to 2.7 g/cm³, does not have sufficient buoyancy to drive upward diking in a compressive crust of only slightly greater density (e.g., Loucks, 2021), and this mechanism for controlling basalt eruptions may be represented in the spatial patterns. These extrusive patterns are possibly influenced by melting loci in the upper mantle and by many other factors that are beyond the scope of this study.
The Black Mountain field is unique in the central Mojave Desert with its much greater areal extent and volume. Basalt is exposed across two mountain tops (Black Mountain and Onyx Peak) as well as west and southwest in lowlands (Miller et al., 2020) that slope toward Harper Lake (Fig. 1). East of the mountain exposures, basalt of identical chemistry was encountered in boreholes in the Superior Lakes area (Buesch, 2018), and southwest of the exposed basalt, boreholes have encountered the basalt over a wide area (Fig. 1). A few boreholes near Harper Lake (Fig. 1) encountered basalt 40–60 m thick, from which we infer that lava was ponded in a physiographic low. We infer a minimum areal extent of 630 km² for this field, which is larger than the Cima field and over 4 times larger than the next largest field, Amboy. Its thickness indicates an eruptive volume of approximately 19 km³, which dwarfs other basalt fields of the Mojave Desert. Dibblee (1968) inferred vents along the Harper Lake fault cutting the field, based on greater thickness of flows near the fault. Glazner and Bartley (1994) noted that features similar to vent areas exist on Black Mountain. Much remains to be learned about this large field. Compositionally, the field is basalt to basaltic andesite.

Most, if not all, young basalts in the Mojave Desert are olivine-bearing, and many contain plagioclase and augite. Comprehensive geochemical data are published for only a few basalt fields, but available data indicate that most lavas other than those at Cima field range from basanites through basalts to basaltic andesite composition. These trends occur both within fields and among all the fields. Neville et al. (1985) noted that the older xenolith-bearing basalts in the Ruby Mountain and Fry Mountain fields are distinguished from younger xenolith-bearing basalts in the central Mojave Desert by having lower SiO₂ (basanite) and higher TiO₂ and MgO. However, more extensive chemistry for the xenolith-bearing fields shows that the 2.1-Ma xenolith-bearing basalt at Dish Hill and the undated Deadman field are similar to that at the Fry and Ruby mountains, with all having low SiO₂ values between 42.8 and 45.3% (basanite) and similar TiO₂ and MgO as well. Thus, location is not a control on basalt chemistry among the xenolith-bearing basalts, and basanites have erupted at two distinct times. The Cima field is distinct from other xenolith-bearing and nonxenolith-bearing basalts of the Mojave Desert, with a trend from basalt through trachybasalt to basaltic trachyandesite. Its lavas vary compositionally with age and the associated chemistry is consistent with mid-ocean rise basalts produced from oceanic lithospheric mantle that replaced previous continental mantle (Farmer et al., 1995).

Discussion

The temporal pulses of basaltic volcanism in the Mojave Desert we have outlined may be caused by or associated with tectonic events that change the ability of magmas to rise through the crust and by mantle perturbations such as periods of enhanced heat flux and greater shear rates, among many other possible causes. It has long been stated that distinct periods of activity along the San Andreas fault zone may drive tectonic changes in the ECSZ. One example is the Pliocene initiation of the Big Bend as a driver for compression across strike slip faults in the northern Mojave Desert (e.g., Yule and Spotilla, 2010). As we gain knowledge about the pre-Holocene behavior for ECSZ faults, patterns may emerge. For instance, Nuriel et al. (2019) noted that dates on ECSZ fault-zone opal deposits may be interpreted as increased fault activity starting about 2 Ma and peaking at ~1 Ma, perhaps coincident with the youngest basalt pulse (Table 1). This timing is similar to that inferred for much of the uplift and thrusting of the San Bernardino Mountains blocks (e.g., Yule and Spotilla, 2010). Older pulses of basaltic volcanism may show similar correlations with tectonic activity. Loucks (2021) argued that basalt chemistry in continental arcs is associated with the stress state in the upper lithosphere, with increase of SiO₂ from 50–55% (basalt to basaltic andesite) for non-compressed lithosphere to 57–65% (andesite) for compressed lithosphere. No studies have established whether similar chemistry changes may hold for changes in compression in thin, high-heat-flow crust such as that of the Mojave Desert. The young lavas of the Mojave are mostly basanite and basalt, but three fields (Cima, Black Mountain, and Bicycle Lake basalt fields) are presently known to be higher silica; basaltic trachyandesite (Cima) and basaltic andesite (Black Mountain and Bicycle Lake). These fields with basaltic andesite and basaltic trachyandesite are <5, 3.9, and 4.5 Ma, respectively; at this time, it is unclear if they stem from periods of greater compression or have other origins.

Using the spatial data (Fig. 1) and other published information on the basalt fields, we address below a few interesting questions about the fields in the Mojave Desert.

How did these fields form and what parts persist? Many of the basalt fields include one or more lava fields and cones, but some fields have no identifiable vents, which might result from enhanced erosion of that part of the field. Cones (including poorly preserved remnants) are preserved in ~60% of the fields, and fields with no cones (including one field with a possible vent structure) form the rest. Remnants of all the fields have persisted for up to 12 my., but significant amounts of 12 fields (including cones with well to moderately preserved morphology) are preserved in fields younger than 6 Ma. Although fields older than 6 Ma typically only have lava fields preserved, two exceptions are the ~8 Ma Pioneertown field and the 7.5 Ma part of the Cima field. Each field has experienced significant erosion, but degraded cones are still identifiable. The preservation of cones and lava flows depends on the characteristics of these features and how they formed (Valentine, 2022), and the intensity of erosion or structural disruption by faulting and folding.

Where cones are preserved, they consist of scoria and spatter deposits resulting from explosive eruptions of
fragmenting lava, and some cones are partially mantled by spill-over lava flows. For the fields in this study, there are no well-documented descriptions of phreatomagmatic eruption deposits or maars, although we suspect that several maars existed in the Cima field based on analysis of satellite imagery (Buesch and Hook, 2022, Phil Gans and Greg Valentine written communications, 2021), and preliminary mapping in the Amboy and Ash Hill fields. Phreatomagmatic landforms tabulated in a regional study of Quaternary basalt fields of the southwestern U.S. form ~10 percent of the volcano types (Valentine et al., 2021), which might explain why maars have been overlooked in the young basalt fields of the Mojave Desert. Each of the cone components (scoria to agglutinate to lava mantle) is progressively more resistant to erosion, although the weakest component is typically the critical factor in long-term preservation. Vents such as cones can have lava flows emanate from the crater, or from the basal margins of the cone, and each process can modify the shape of the preserved cone and redistribute parts of the cone as rafts on lava flows. Nearly all cones we have studied in the field or by aerial imagery have a notable component of spatter, agglutinate, and short spillover lava flows, which probably contribute to their preservation. Careful studies of cone morphology (e.g., Zarazuza-Carbajal and De la Cruz-Reyna, 2021) find correlations between morphology and age for cones <50 ka. This approach could be extended to older cones using Mojave Desert examples.

Effusive eruptions produce long fissure or point-source vents that have proximal features such as spatter or agglutinate deposits and ramps. In the basalt fields we compiled, such fissure vents have not been identified. In the Pisgah and Amboy fields, effusive vents are associated with (1) elevated ponded lava or lava lakes, or (2) dome or pyramidal shaped structures formed by injection of lava that uplifted and deformed older flows. Because these vent structures are primarily lava flows, they can be very resistant to erosion, and recognizable from subtle topography. Most of the lava fields are formed by pāhōehoe, ‘ā‘ā, or block lava, and in several fields pāhōehoe flows typically have a longer runout compared to ‘ā‘ā. The distribution of the flows results from flow across the ground surface, advancing the flow front, and can be enhanced by formation of lava tubes that deliver lava to the advancing flow front. In low-relief areas, lava tubes can form but remain filled as the lava flows toward the advancing front; in areas with greater ground slope the upstream lava can flow out, leaving an open tube and collapsed skylights into the tube.

Do the basalt fields tend to form in a particular landscape position? The young fields in Figure 1 lie in valley bottoms (e.g., Amboy), high on gently sloping pediments (e.g., Cima), on bedrock and mountain-front alluvial fans (e.g., Pioneertown), and in steep mountainous terrain (e.g., Pipkin and Deadman Lake). The large fields, such as Black Mountain, cover a wide variety of terrain. A few fields include dikes through Miocene and (or) Pliocene alluvial conglomerate (e.g., Deadman Lake, Howard et al., 2013), and dike and funnel-like feeder vents in granite and gneiss (e.g., Ruby Mountain, Neville and Chambers, 1982) that indicate extrusive vent areas in mountains. Although vent areas are not known for several fields, it appears that local relief is not a factor for eruption location.

How long do fields persist? Although the probability that a basalt field is preserved is much greater for young fields than old fields, we suspect that we have cataloged nearly all young fields for the Mojave Desert for the following reasons: 1) Regional geologic mapping of the past century identified all flat-lying basalts as Quaternary, and we have examined all of these to find that many were much older than our 12–0 Ma timeframe. The oldest are ~19 Ma and several in western Joshua Tree National Park date at ~15 Ma. From this evidence, it appears that basalt flows are extremely persistent in the landscape unless buried in valleys, as they are at Fort Irwin (Bicycle Lake field), Harper Lake and Superior Lake (Black Mountain field), and at Dish Hill, or where they are segmented and deformed by tectonism. 2) The younger intact fields of the Mojave Desert range in location across highlands and lowlands and at least parts of these will be spared from burial. 3) Although the Mojave Desert is tectonically active, deep basins are not forming. The major faults are strike-slip and do not create as much relief as faults with greater components of dip-slip offset. For these reasons we consider our data set to be complete. Basalt fields may persist for over ten million years on the surface of the Mojave Desert.

Will there be another eruption soon? The fairly recent eruption at the Cima field (12 ka) could indicate an eruption hazard for the Mojave Desert. An evaluation of volcanic hazards for the state of California indicates no hazard (California has active and hazardous volcanoes (usgs.gov)) for the part of the Mojave Desert we studied, but finds potential for volcanic hazards in the nearby Salton Buttes area, Ubehebe Crater, and Coso volcanic field. It is difficult to make predictions for basaltic eruptions that are sparse across the landscape. Firstly, magma is typically generated from 30–40 km deep, and it can reach the ground surface in ~10 hours because it is not necessarily stored in a shallow magma body. As a result, impending eruptions can be difficult to detect by tracking magma movement although a propagating dike that can be monitored seismically. Secondly, basaltic eruptions are more readily predicted if they occur in an eruptive complex in which numerous past eruptive products are available for determining age relations and average duration between eruptions. Our compilation indicates that the Cima field is the only Mojave basalt field with spatial and temporal overlaps for eruptions. With improved age control, an estimate of time to the next eruption might be possible for that field. All other basalt fields are widely separated in space and time, seemingly indicating that there is an exceedingly small chance of an eruption in our lifetime.
Concluding remarks
This compilation uncovered new spatial and temporal patterns for basalt eruptions and pointed out basalt fields that have been little studied. The promise for significant breakthroughs with future comprehensive study of these young basalts is great. We hope that the compilation and the ideas expressed in this paper will serve as a springboard for future studies.

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The Dish Hill volcanic complex, San Bernardino County, California

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ABSTRACT—Dish Hill, also historically referred to as Siberia Crater, is located north of National Trails Highway (Route 66) in the Mojave Desert 20.3 km (12.5 miles) west of Amboy, California. K-Ar ages on amphibole in ultramafic xenoliths and fission-track ages from granitoid xenoliths that range from 1.9 to 2.1 Ma indicate that it is one of the oldest volcanoes in what has been referred to as the Amboy-Barstow Axis. The three volcanic edifices that form the complex are Dish Hill (Siberia Crater), Hill 1933, and Hill 1068. All three volcanoes erupted basanite lava that formed breached cinder cones built by layers of cinders, blocks, bombs, and spatter. The spatter strengthened the cones, which allowed them to fill and spill over the rim before being breached. Lava flows from Dish Hill and Hill 1933 traveled approximately 2.5 km (1.5 miles). Hill 1068 is south of National Trials Highway and protrudes out of the alluvium. Dish Hill developed on Mesozoic granite and the lava flows and pyroclasts contain mantle-derived xenoliths with olivine, clinoxyroxene, orthopyroxene, and kaersutite that are visible in hand specimen. Abundant xenoliths are located in pyroclastic blocks and bombs that weather out of the volcanic breccia. The presence of zeolites formed in the groundmass of lava flows and cone building spatter are indicators that the magma was water rich. Dish Hill rocks and processes have been analyzed in numerous papers on petrologic development of mantle melts, characterization of mantle and lower lithospheric xenoliths and inclusions in the magmas, and the tectonic evolution of the Mojave lithosphere.

Introduction
Southeast of Barstow, a belt of Pliocene to Quaternary olivine basalt volcanoes can be accessed from Interstate 40 and National Trails Highway (Route 66). Included in this belt is a complex of three volcanoes commonly referred to as Dish Hill, and herein named the Dish Hill complex. Brady and Webb (1943) refer to this cluster as Siberia Crater whereas Wilshire and Trask (1971) refer to a small cone to the south of Dish Hill volcano as Siberia Crater. Recent literature refers to the complex of three separate cones as Dish Hill in regional maps, but the individual volcanoes are Dish Hill, Hill 1068, and Hill 1933 (Wise, 1966; Figure 1).

The three volcanoes in the Dish Hill cluster erupted basanite, the name given to a group of basaltic rocks composed of calcium-rich plagioclase, clinoxyroxene, feldspatoids, and olivine; geochemically, basanite has low (41–49 wt%) SiO$_2$ and moderate alkalis (3–9 wt%) Na$_2$O+K$_2$O (Le Maitre, 1989). Basanite lava in the Mojave Desert occurs only at Dish Hill, Deadman Lake, Lead Mountain, Pisyah, Sunshine, Lavic, and Pipkin (Wise, 1969). Dish Hill and Hill 1933 were deposited on Jurassic and Cretaceous granitic rocks whereas Hill 1068 appears to be partly buried by alluvium (Figure 1). Each volcano has cones built by layers of cinders, bombs, lapilli, blocks, and lava spatter that fused together while hot and viscous. It is plausible that lava spatter strengthened Dish Hill and Hill 1933, thus allowing lava to fill and spill over the rims. Each volcano has lava flows erupted from spillover of crater rims, fissures on the sides or near the base of the cone, or the central vent after the cone was breached. Based on the morphology of the cones and exposures of dikes in the cone, Wise (1966) estimated as much as 30 m of loose material has been eroded from the cone surfaces.

Dish Hill is early Pleistocene in age based on three dates: a 2.1 ± 0.2 Ma fission-track age determined on apatite in a granitic inclusion (C.W. Naeser in Wilshire and Trask, 1971), a 2.03 ± 0.12 Ma K-Ar age on a selvage (fracture-fill) amphibole (M.A. Lanphere in Wilshire et al., 1980), and a 1.9 Ma K-Ar date on amphibole (Wilshire and Nielson-Pike, 1986). Dish Hill is one of several young basaltic volcanoes in the Mojave Desert that has a variety of ultramafic xenoliths from the upper mantle rocks and deep crustal granitoid rocks that were brought from those depths by basanite. These basanites appear to have crystallized minor amounts of zeolite in the groundmass (Wise, 1966). These rocks and processes have been analyzed in numerous papers on petrologic development of mantle melts, characterization of mantle and lower lithospheric xenoliths and inclusion in the magmas, and the tectonic evolution of the Mojave lithosphere (Wise, 1966; Wilshire and Trask, 1971; Wilshire et al., 1980; Wilshire and Nielson-Pike, 1986; Wilshire et al., 1988; Luffi et al., 2009).

Centers of volcanic activity
Hill 1933 is northeast of Dish Hill and consists of a scoria cone with basanite lava flows on several sides (Wise, 1966; Wilshire and Nielson-Pike, 1986; Figure 1). Flows on the west and south sides of the cone appear to have emerged from a series of dikes. The final event was the breaching of
Hill 1933 with lava flowing northeast and then following the drainage towards the southeast for approximately 2.5 km (1.5 miles). This flow rafted parts of the cone.

Hill 1068 is located south of National Trails Highway; it protrudes out of the alluvium approximately 52 m (170 ft), and was breached toward the northeast (Wise, 1966; Wilshire and Nielson-Pike, 1986). Evidence of flow activity is covered by the alluvium. The cone is similar in nature and composition to the other two volcanoes.

Dish Hill is a scoria cone with five associated flows, and dikes that locally cut the scoria cone might have fed small flank effusive vents (Wise, 1966; Wilshire and Nielson-Pike, 1986; Figure 2). The scoria cone has upper pyroclastic deposits referred to as agglutinate where clasts are sintered into a hardened rock, and the lower slopes of the cone are volcanic (pyroclastic fallout) breccia with lapilli to blocks clasts with various amounts of tuffaceous matrix. There are abundant xenoliths in the bombs and blocks (Figure 3). Numerous cavities, especially in the volcanic breccia, form where the bombs and blocks weather out of the breccia, or by variable matrix strength due to the alteration of the ash to palagonite (Figure 4).

Basanite flows are located northeast, southeast, northwest, and southwest of Dish Hill. The lava flow in the southeast is only a small remnant deposited on the granitoid bedrock. Lava flows to the northeast and part of the lava flows to the southwest appear to have spilled over the crater rim prior to a breaching event. The largest flow, on the northwest side of the cone, breached the cone and followed a drainage towards the southwest for flow length of about 1.6 km (1 mile). This flow is capped by large amounts of agglutinate that are probably rafted cone fragments (Wise, 1966).

The lava flows at Dish Hill (and Hill 1933) were relatively fluid, are up to 7.6 m (25 ft) thick, and have 0.7–1.7 m (2–5 ft) thick vesicular zones at the top and bottom of each flow, with a sparsely vesicular core. All flows contain xenoliths (Wise, 1966). At Dish Hill, the northeast lava flow is dominated by titanaugite xenocrysts (up to 25 mm) whereas the southern flow contains kaersutite (up to 35 mm) and abundant peridotite xenoliths (Wise, 1966). The tops and bottoms of the flows have up to 20 percent glass resulting from rapid chilling. The crystallized lava and pyroclasts are extremely fine grained (<0.1 mm) with microphenocrysts (<0.5 mm) of olivine, and groundmass grains are olivine, titanaugite, titanmagnetite, ilmenite,
and small plagioclase. In the cores of lava flows, there are small patches of phillipsite–chabazite intergrowth that Wise (1966) interpreted as initial crystallization of the glass rather than feldspathoids. Wise (1966) proposed that zeolites crystallized during initial cooling of the lava. Rapid cooling, after eruption of the lava, resulted in a very fine-grained texture. Spatter-built cones are a good indicator of water rich magma. It is likely that olivine (the microphenocrysts) started to crystallize in a deep-seated chamber just before or at the beginning of the eruption. Once lava reached the surface, the groundmass crystallized rapidly resulting in an enrichment of alkali rich water and formation of zeolites.

Fig. 2. Geologic map of Dish Hill and vicinity from Wilshire and Nielson-Pike (1986).

Fig. 3. Volcanic bomb with xenocryst of kaersutite and peridotite xenolith. The dark mineral in the NW corner of the picture is kaersutite. The peridotite is left of the quarter, and contains olivine (olive green), clinopyroxene (apple green), orthopyroxene (dark brown), and kaersutite (glassy black). The quarter is 2.5 cm (1.0 in). Photo by Bruce W Bridenbecker.
Petrology of the mantle xenoliths

Dish Hill lavas have ultramafic xenoliths and phenocrysts derived from the xenoliths. Figure 3 displays xenoliths of mantle minerals including olivine, clinopyroxene, orthopyroxene, and kaersutite. It is a classic location for xenoliths because of the range in composition of xenoliths and minerals that represent rocks from various depths (pressures and temperatures) in the mantle, and textural studies provide insights to structural deformation.

- **Xenoliths** — Each type (name) of a xenolith implies a composition (especially the ultramafic rocks that have different ratios of olivine, orthopyroxene, and clinopyroxene), and these rocks can have a specific mineral (such as Cr-diopside) as a defining adjective. Xenoliths include: peridotite and minor amounts of partially fused granite (Wise, 1966), and peridotite, lherzolite, wehrlite, harzburgite, pyroxenite and partially fused granite (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009).

- **Minerals** — Minerals as xenocrysts and fracture filling (selvages) can have variations of minor or trace elements that have petrologic implications (details not included). Xenocrysts include titanaugite and kaersutite (Wise, 1966), and kaersutite (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009). Fracture fillings include kaersutite, phlogopite (mica), rareapatite and opaque oxides, and very rare plagiooclase (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009).

- **Structural features** — Structural features in the xenoliths include granoblastic texture, kink banding of olivine, foliation from mylonitization and recrystallization (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009). In addition, Wilshire and Trask (1971) noted: (1) sharp contacts of pyroxene-rich and olivine-rich rocks, (2) polished fractures (by fusion of amphibole rather than abrasion polishing), (3) as many as six faceted surfaces on a xenolith (the breaking of much larger blocks), and (4) veins of different compositions that cross or offset foliation or kink bands in olivine (Wilshire and Trask, 1971).

- **Geochemical analyses** — Geochemical analyses published (or referred to) in papers includes major oxides, minor-trace elements, and Rb, Sr, H, O isotopes (Wilshire and Trask, 1971; Wilshire et al., 1980; Luffi et al., 2009).

- **Models for magma melts and xenolith inclusion** — Mantle or lower lithospheric rocks that were partially melted or were metasomatically altered and then included as xenoliths at different pressures and temperatures indicate complex processes. One trend is that the xenoliths were included at shallower and lower temperatures as the field evolved. To explain these processes, Wilshire et al. (1980 and 1988) invoked a rising diapir model, and Luffi et al. (2009) invoked sampling of tectonically subcreted and imbricated oceanic lithosphere emplaced during low-angle subduction.

**Summary**

Dish Hill and two nearby volcanoes provide good examples of the development of pyroclastic and agglutinate cones and lava flows from relatively low viscosity basanite magma. The lavas are unusual in having zeolites as groundmass minerals. Mantle-derived xenoliths and xenocrysts occur in lava flows and many pyroclastic bombs, and the minerals and textures in xenoliths provide a complicated history of the upper mantle.

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**References**


Aligned feeder dikes for the Deadman Lake volcanic field

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The Deadman Lake and Dish Hill clusters of Pleistocene basaltic cinder-agglutinate cones and associated small flows in the Mojave Desert are known for their mantle-derived ultramafic xenoliths (Wilshire and Trask, 1971; Wilshire et al., 1980, 1988; Wilshire and Nielson-Pike, 1986; xenoliths are lacking in the nearby younger late Pleistocene Amboy and Lead Mountain basalt flows; Fig. 1). The basalt at Dish Hill was dated to about 2 Ma by K-Ar and fission-track dating (Wilshire et al., 1980; Wilshire and Trask, 1981), which suggests that the Deadman Lake and Dish Hill clusters are early Pleistocene. These two clusters flank the south and north sides respectively of an alluviated valley that forms part of the Barstow-Bristol trough. The Deadman Lake cluster contains remnants of 12 or 13 cones (Wilshire et al., 1988), whereas Dish Hill is one of 3 cone remnants clustered on the north side of the trough. Small-wavelength aeromagnetic highs in the alluviated valley between the Deadman Lake and Dish Hill clusters likely signal several concealed basaltic vents between and connecting them. Wilshire et al. (1988) stated that the Deadman Lake cones are aligned on a northeast trend. Several basaltic dikes associated with these cones strike northeast at 026°–063° (e.g. Schafer et al., 1950; Howard et al., 2013). This strike and similarly aligned fissures (averaging about 040° azimuth) in Broadwell Lake playa, 30–40 km to the northwest (Dibblee, 1967), are consistent with extensile fracturing aligned along a regional north-south to northeast-southwest greatest horizontal compressive stress (Zoback and Zoback, 1980). The Barstow-Bristol trough, a parallel Lucerne-Dale trough 40–50 km to the south, and the intervening Bullion Mountains highland form a series of topographic wave forms approximately perpendicular to this stress direction. The broad wave forms may record long-wavelength buckling of the upper crust (Howard and Miller, 1992; Howard and Cox, 2001).

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Mohave Nation version of the formation of Amboy Crater

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Introduction

Native American people have lived in the Mojave Desert for many thousands of years. Like many civilizations, they would tell stories around the evening fire to the younger generation as a way of explaining many of the natural features they observed. Such is the case with the Mohave Nation story about the formation of Amboy Crater. AH MOTT KAH PEE THOYAH was a hunter who used a peculiar hunting technique. Rather than summarize the story, I present the narration as found in the San Bernardino County Museum Association publication, “That All May Know—The Lore of The Mohave Nation.”

AH MOTT KAH PEE THOYAH

A long time ago the people had various ways of making their living. Some of them fished, some were hunters, etc. There was a hunter by the name of AH MOTT KAH PEE THOYAH (he can go down under the earth). That is related to the way he hunted for a living. He hunted birds and rabbits and he used fire in the strange method he pursued his hunting. He would go to any thick forest or brush and would build a fire in a circle all around the thick brush or trees—carefully built so no animals could escape from the fiery trap.

Then he went to the center of the place that the fire was encircling and sang a song that caused him to sink into the earth. Here is his song: AH MOTT KAH PEE THOYAH, KAH PEE THOYAH, EEHAH VEE ROPES, AH MOTT THEE THON EETHON EE EE. This means “When my name is sung, I go farther and farther down into the ground.” When the song was finished the first time he sank down into the ground above his knees. Then he repeated the song four times and he was out of sight under the ground. When he thought the fire had burned itself out, he then came up out of the ground and he would find lots of rabbits, deer, mountain sheep, etc., that had stayed in the brush that were cooked and ready to eat. He would gather all that he desired and then be on his way home. After he had used up all the meat from the last kill, he would then start out and find a new location to hunt. That was his method of hunting.

One time after he had performed his hunting process and had come back up out of the ground, he found a Fox there looking over all the dead animals. AH MOTT KAH PEE THOYAH said, “You may have all you want, Mr. Fox. I'm giving you all you want.” The fox shook his head and said “No! That stuff stinks. I don't eat such smelly stuff. I hunt and I kill my own way.” Then the Fox thanked the Fire Hunter and left, but, before he left, he asked, “What do you sing and how do you hunt with fire?”

AH MOTT KAH PEE THOYAH told him that he simply started a fire around himself in a complete circle and then sang a song four times and that caused him to go under the ground so he wouldn’t get burned. The Fox said, “Just for curiosity’s sake, I’d like to have you demonstrate for me.” So, to accommodate him the hunter sang [the] song for him, and the Fox memorized the song and went on his way without taking any of the food he had been offered.

The next day the fox decided that now since he knew the secret to the Fire Hunter’s success, he would hunt the same way. So, he picked out as his location to hunt the land just south of Amboy, California, where the volcanic crater is located today. He worked fast, and got the fire thoroughly started all around him, then he went in to the center of the ring of fire and began to sing the song that AH MOTT KAH PEE THOYAH had taught him the day before. But he didn’t go down a single inch into the ground. In desperation, he repeated the song four times and still nothing happened. In death he was still making a tremendous effort to send himself down into the ground just like the Fire Hunter had done, for just as you have
probably surmised, the fire had gotten out of control and cremated the Fox on the spot.

But, the fire didn’t stop. It kept burning more and more and soon the earth itself started to burn. Then the rocks began to burn and the rocks melted and flowed over the desert like a stream of boiling tar. The melted rock kept running until it reached Newberry, and it didn’t stop there. It even kept on running until it reached Barstow.

AH MOTT KAH PEE THOYAH came along and saw what had happened and he stopped the fire. When the fire had subsided, there was a great big crater left and is still there today. Then the Fire Hunter changed himself into a little worm and he is in this form yet today.

Whenever you happen to be around sand hills or around any sandy place, you can easily find the marks on top of the sand where he travels around under the sand. Also, you will find little crater-shaped holes where he gets down under the ground looking for insects. Now he still goes down— not head first—but backwards into the sand, just as he did when he was a man.

This is a very true story, but if you don’t believe it, just look around sandy places and you can see the tiny craters. Then try to find the little worm AH MOTT KAH PEE THOYAH. He is there but you won’t find him for he is already buried under the earth and is laughing at you.

Airborne infrared spectral mapping of the southern Marble Mountains, San Bernardino County, California

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ABSTRACT—An airborne survey of the southern Marble Mountains (San Bernardino County, California) was conducted with a wide-swath longwave-infrared spectral imager. The longwave-infrared is the canonical molecular fingerprint region where the diagnostic spectral features of most molecular materials reside. Consequently, it has been used extensively for identifying and discriminating the major classes of rock-forming minerals (silicates, carbonates, sulfates). Laboratory measurements of ground-truthed field samples provided insight into the spectral traits of the various lithologies encountered and were used to guide the selection of wavelengths used in constructing the imagery-derived maps. The high spatiotemporal resolution of the imagery used for this study enabled clear delineation of changes in lithology and the contacts between different formations. Despite difficulties in definitively separating compositionally similar lithologies, the derived map corresponded closely to previous geologic maps produced of the area, while also identifying several errors in the earliest map. The high spatial resolution (2 m) of the imagery allows lithologies and contacts to be mapped with much greater fidelity over much larger areas than is practical through conventional field mapping techniques, while the large-area coverage permits mapping to be extended into areas that are difficult or impossible to access on the ground. The investigation demonstrated that high-resolution longwave-infrared spectral imaging offers significant potential for aiding the mapping of poorly understood horizons and co-validating fieldwork-derived maps.

Introduction
The Marble Mountains are located in San Bernardino County, CA about 150 km east of Barstow, between Interstate 40 and Route 66. They are well known for well-preserved Lower Cambrian olenelloid trilobites in the Latham Shale, and the area south of Route 66 has been a popular site for fossil collectors. The area north of Route 66 has been given protective status as the Trilobite Wilderness by the Bureau of Land Management. The trilobite biostratigraphy in the area has been presented in Foster (2011) and Mount (1973, 1974). The southern half of the Marble Mountains was chosen for a remote sensing mapping exercise because of well exposed rocks and easy access to outcrops, its simple stratigraphy and existence of published geological maps for comparison.

Southernmost Marble Mountains geology
The geology of the Marble Mountains was mapped by Kilian (1964) and the area south of Route 66 was more recently mapped by Foster (2011) and Kenney (2011). The Cambrian stratigraphic section in the southernmost Marble Mountains is presented in Figure 1. About 350 m of Early to Middle Cambrian sediments were deposited on Precambrian crystalline basement which Kilian (1964) divided into three main types, in decreasing age: gray coarse-grained granite, gray diorite and porphyritic red granite. In the southernmost Marble Mountains, the basement is mainly the red coarse-grained granite which contains K-feldspar phenocrysts to 3 cm with quartz, plagioclase and biotite.

In increasing age from Lower to Middle Cambrian, the sedimentary units, summarized from Foster (2011) are:

Wood Canyon Formation: 110 m of cross-bedded gray quartz arenite, deposited on Precambrian granite basement, that weathers reddish brown to dark gray as a result of desert varnish. 7 m above the granite contact is a 2 m thick bed of coarse-grained conglomerate containing 1-3 cm wide milky quartz clasts.

Zabriskie Quartzite: 36 m of gray medium-grained quartz arenite that weathers tan to golden brown from desert varnish.

Latham Shale: 19 m of greenish gray to reddish brown fissile shale that locally contains abundant olenelloid trilobite fossils (Mount, 1973, 1974).

Chambless Limestone: 48 m of dark gray oncoid-bearing limestone that forms resistant ridges.

Cadiz Formation: 120 m of interbedded green-gray to reddish-brown shale and sandstone, reddish-brown dolomite, oolitic limestone and dolomitized limestone.

Bonanza King Formation: Only partially exposed in the southeastern tip of the southernmost Marble
Mountains and consists of gray to reddish brown dolomitic limestone.

Kilian (1964) mapped Tertiary volcanic rocks, ranging from basalts to tuffs overlying much of the Paleozoic sedimentary sequence in the central part of the range. These units are Miocene in age and consist of (from oldest to youngest) (1) locally exposed dacite lava flows and tuffaceous deposits, (2) basalt flows, (3) possible thin interval of sedimentary rocks, (4) Peach Spring Tuff, (5) possible thin interval of sedimentary rocks, and (6) locally exposed basalt flows (Buesch and Harvey, 2022).

**Infrared spectroscopy**

Above absolute zero, the interatomic bonds in molecules stretch, bend and rotate at frequencies that depend on the mass of the atoms, interatomic distances, and bond strengths. The resulting vibrations give rise to characteristic spectral transitions diagnostic of chemical functional groups within the material in question. At ambient temperature, the frequency of the molecular vibrations falls in the so-called thermal-infrared wavelength range (2-25 μm). This is the basis of infrared spectroscopy, whereby materials probed with infrared light impose their own characteristic spectral structure on the transmitted or reflected light and the result is analyzed to determine the chemical composition of materials. The infrared transmission spectrum has been used in this way for decades by organic chemists and reference libraries containing > 100,000 spectra have been compiled for the identification of materials. The technique has been less widely used for inorganic materials and minerals, where generally only the anion groups produce absorption bands, with different cations tending to shift the wavelength of the anion absorptions. The RRUFF Project has endeavored to increase the number of reference infrared spectra of minerals (Lafuente et al., 2015).

In contrast to transmission spectra, which have transmission minima (absorption bands), reflectance spectra have reflectance maxima (reststrahlen bands) at approximately the same wavelengths. Reflectance infrared spectra have been less widely used for material identification because specular reflectance spectra from mirror-like surfaces differ from diffuse reflectance spectra from rough surfaces, and samples are rarely solely specular or solely diffuse. Furthermore, diffuse reflectance spectra show variations which are dependent on particle size or surface roughness, as the contribution from volume scattering over surface scattering increases with decreasing particle size (Hunt and Vincent, 1968; Salisbury and Wald, 1992; Mustard and Hays, 1997). In general, spectral contrast and intensity decreases with decreasing particle size and the intensity from the broad transparency feature at longer wavelength increases with respect to the reststrahlen band (Salisbury and Wald, 1992). Figure 2 compares the transmission, specular and diffuse reflectance spectra of quartz.

Above absolute zero, matter radiates electromagnetic energy that at ambient temperatures is in the infrared wavelength range. This emitted radiation can interact with materials by the aforementioned processes and produce emissivity minima associated with the reststrahlen bands. This can be exploited by remote sensing methods since no interaction with the material is required but libraries of reference emittance spectra are required to make identifications. Laboratory measurements of emissivity are difficult to make and typically reflectance measurements are made instead. Diffuse reflectance (R) can be related to emissivity (ε) by Kirchhoff’s Law: ε = 1-R (Hunt and Vincent, 1968; Salisbury et al., 1994). The fact that diffuse

![Figure 1. Stratigraphic section of Cambrian sedimentary rocks from the southernmost Marble Mountains, from Foster (2011).](image-url)
Reflectance spectra vary with particle size complicates remote sensing identifications since multiple spectra, representing various particle sizes, must be included in spectral libraries.

**Laboratory measurements**

Laboratory Fourier transform infrared (FTIR) measurements were made at 4 cm⁻¹ resolution with a dry-nitrogen-purged Thermo Scientific Nicolet model 6700 FTIR spectrometer equipped with DTGS and MCT-A detectors. Biconical diffuse reflectance spectra were measured from the upward-facing surfaces of rock samples. These spectra are from an area of ~ 1 mm², and were performed using a Harrick "praying mantis" accessory after rock samples were trimmed to less than 2 cm × 2 cm × 1 cm, in order to fit in the attachment. Labsphere Infragold was used as the background reference. Biconical reflectance spectra reproduce spectral shapes accurately but are not quantitative. This is usually sufficient for identification purposes. Hemispherical reflectance measurements are quantitative but have significantly poorer signal to noise and require very long acquisition times, which may not be practical for large numbers of samples.

Laboratory measurements of diffuse reflectance are generally used to record reference spectra for remote sensing libraries, but they sample a very limited area, ~1 mm dia for biconical and 5 mm dia for hemispherical. Many rocks have crystals of that size or larger so that in order to record representative spectra, the rocks are ground to particle sizes of 75 µm to 250 µm and < 75 µm which homogenizes the sample on the scale of the sampling volume. The size fraction < 75 µm, which is actually dominated by particles much less than 75 µm, produces spectra representative of fine-grained material, while the 75–250 µm fraction produces spectra representative of coarse-grained material. Most spectra in reference libraries, therefore, are from fresh fracture surfaces which are not representative of natural rock surfaces in the field. The latter show the effect of chemical weathering, which changes the composition of the surface, and physical weathering, which can change the surface roughness (Kirkland et al., 2002, 2003). Furthermore, surface coatings such as desert varnish or caliche may form, which partially or totally mask the underlying material. For this reason, we performed a ground truth survey of the southernmost Marble Mountains in October 2013 to collect samples of representative rock types, noting which surface was up-facing. Biconical reflectance measurements were made from the samples in order to add the reference spectra to the Aerospace Solids Library and to gain insight into the spectral features which would allow greater understanding of the remote sensing images.

Figure 3 shows the diffuse reflectance spectra of the main sedimentary formations in the southernmost Marble Mountains. The reststrahlen bands between 8.0 µm and 10.0 µm are from Si-O bonds in silicates (quartz and clays) while the bands at 6.45 µm and 11.26 µm are from the CO₃ groups in the carbonate facies. The carbonate band at 6.45 µm is outside of the Mako sensor spectral range. It is noted that the spectrum of the Chambless limestone is relatively pure calcite while that of the sandstone member of the Cadiz Formation is a mixture of calcite/dolomite and quartz. The spectrum of the Latham shale is dominated by the clay band at 9.62 µm. Figure 4 shows the diffuse reflectance spectra from several of the igneous rocks that occur in the southernmost Marble Mountains. It is noted that the spectrum of the Precambrian granite is not representative because the grain size greatly exceeds the diffuse reflectance spot size. The spectra are...
dominated by reststrahlen bands of silicate minerals (quartz, feldspars, micas, pyroxenes, amphiboles) between 8.0 µm and 10.0 µm. Since the majority of these rocks are heavily varnished, the clay band at 9.62 µm is also very prominent.

Desert varnish consists of thin coatings (several to 10's of µm) of predominantly clays and iron and manganese oxides that form very slowly over time (µm's/1000 yrs) on intermittent moist rock surfaces which trap very fine windborne particles, i.e. clays (Potter and Rossman, 1979; Dorn and Oberlander, 1981). An example is shown in Figure 5. Since the thickness of the varnish increases with time, and shows micro-laminations, it has been suggested as a way to date petroglyphs (Dorn and Whitley, 1984) and serve as a record of regional climatic conditions (Liu and Broecker, 2013). The thickness of the varnish is on the order of, or substantially thinner than, infrared wavelengths so one would expect to see the underlying substrate in diffuse reflectance spectra. Varnish forms on most rock types in the desert environment except for pure hydrothermal milky quartz veins and carbonates. In the latter, the rock is actually water soluble to a small degree and is not a stable substrate. Figure 6 presents diffuse reflectance spectra of several samples of Wood Canyon...
Formation quartz arenite, with varying thicknesses of varnish, collected from desert pavement. The relatively clean sample (MB1A) shows the characteristic spectrum of quartz which has a broad doublet at 8.31 and 9.20 µm from Si-O bonds with a minimum at 8.64 µm, along with a secondary doublet at 12.50 µm and 12.78 µm. As the varnish thickness increases, the intensity of the quartz band at 9.0 µm decreases, and a peak at 9.62 µm from the silicate group in clay, increases. This variability, depending on varnish thickness, complicates spectral-based mapping of geologic units.

Figure 7 compares the biconical diffuse reflectance of fresh fracture surfaces and varnished surfaces of rhyolite and basalt collected just south of Route 66. It can be seen that there are noticeable differences which may account for remotely sensed spectra not matching spectra in reference libraries. The varnished spectra are dominated by the clay band at 9.62 µm whereas the fresh fracture surfaces have broader bands at shorter wavelength from other silicates. The difference in the two fractured basalt spectra is probably a result of the small FTIR spot size and the presence of plagioclase phenocrysts in the groundmass.

While the growth of desert varnish modifies the reflectance spectrum of a rock by masking the underlying substrate and adding spectral features, physical weathering can produce unexpected and nonintuitive spectra. Mormon Mesa in Clark County, Nevada has significant deposits of terrestrial carbonate hardpan/duricrust (calcrete) that were expected to produce strong calcite spectral signatures, when in fact they were very muted and hard to detect by remote sensing (Kirkland et al., 2002, 2003). This is because the calcrete was very fine grained and when weathered by dissolution, it produces a very microporous texture which produces a greater amount of volume scattering. In contrast, coarse-grained limestone cobbles weathered to produce large faceted grains at the surface, which predominantly produce surface scattering. The effects can be seen in the quantitative hemispherical reflectance spectra where the limestone has much greater intensity (Figure 8).

Remote sensing
The southern half of the Marble Mountains were imaged by the airborne long-wavelength infrared (LWIR) hyperspectral sensor Mako (Warren et al., 2010; Hall et al., 2011) on August 27, 2013. The locations of the two separate data sets are presented in Figure 9. Mako and its predecessor SEBASS have been widely used for...
passive remote sensing and identification of gases and minerals (Tratt et al., 2011; Buckland et al., 2017; Kirkland et al., 2002, 2003; Adams et al., 2017). Mako has also been used to examine the south Bristol Mountains, Old Dad Mountains, and parts of the northwestern Marble Mountains (Buesch and Harvey, 2017; Buesch and Harvey, 2022). These sensors operate in the nominal 7.5–13.5 µm atmospheric window, between strong carbon dioxide and water vapor absorptions. Mako is a 3-axis stabilized whisk-broom system with a swath width programmable up to 110° and 0.55-mrad per pixel field of view with a spectral resolution ~46 nm in 128 channels. The spatial resolution (ground sampling distance-GSD) is determined from the altitude and for this study was 2 m. The Mako instrument suite includes a digital camera that records contemporaneous true-color context imagery of the scene.

Details of the spectroradiometric calibration and data processing provisions for Mako are provided in Buckland et al. (2017). Georeferencing of images is accomplished by the instrument flight software continuously recording position (latitude, longitude, and altitude) from an integrated global positioning system (GPS), the aircraft attitude (roll, pitch, and heading) from the inertial navigation system (INS) and the scan mirror encoder angles. The measured at-sensor radiance includes contributions from the atmospheric transmission and upwelling radiance which impose spectral features (primarily due to H₂O, CO₂, O₃, CH₄, and N₂O) onto the data. Atmospheric compensation is accomplished through the ISAC (In-Scene Atmospheric Compensation) algorithm (Young et al., 2002) which makes use of the natural occurrence of blackbody (or near blackbody) surfaces within the scene (e.g., areas covered by healthy

Figure 7. Biconical diffuse reflectance of fresh fracture surfaces and varnished surfaces of rhyolite and basalt. Spectra normalized to Basalt-fracture MB16B for ease of comparison.

Figure 8. Hemispherical reflectance spectra (common scale) of weathered Mormon Mesa calcrete and Goodsprings limestone.
vegetation or water) and exploits the natural temperature variations over these surfaces to extract transmission and upwelling contributions directly from the measured data.

There are several ways that remote sensing data can be processed and presented. A software toolkit such as ENVI (https://www.l3harrisgeospatial.com/Software-Technology/ENVI) is commonly used. Following atmospheric compensation, a principal component analysis (PCA) is used to identify pixels that have similar characteristics. This analysis can either be run in an automated fashion with no prior knowledge of the geology or guided, when some of the geology is known. The most representative spectrum of a component can then be correlated against a reference library of spectra to obtain an identification. A simpler approach to produce a map, which doesn’t require atmospheric compensation, is to use a method analogous to the normalized ratio approach developed for characterizing forest burn scars in satellite imagery (López-García and Caselles, 1991). In the normalized radiance ratio (NRR) implementation used in this work, for each red-green-blue (RGB) channel, the radiances at two wavelengths (L1, L2) are selected such that L2 correlates with a particular diagnostic spectral band (e.g., silicate) and L1 provides a proximal “background” signal. The NRR is then defined as: (L1–L2)/(L1+L2). A four standard deviation contrast stretch was then performed to convert the image to 8-bit. The RGB images are then combined to produce a color image. Table 1 presents the L1 and L2 wavelengths that were used in creating imagery of the southern Marble Mountains.

The L2 and L1 wavelengths were chosen from a limited subset of wavelengths after several iterations and were able to discriminate different rock lithologies in a definitive manner. The L2 wavelengths represent the minimum in the quartz spectrum (8.63 µm), the longer wavelength feature in the quartz doublet (9.07 µm) and the approximate clay peak (9.51 µm) (see Figure 3). The blue L1 wavelength happens to correspond to the weaker quartz doublet.

Figure 10 presents the NRR map of the southernmost Marble Mountains. The composition of the Wood Canyon Formation and Zabriskie quartzite are so similar that they

Table 1. L1 and L2 wavelengths used in creating imagery.

<table>
<thead>
<tr>
<th>Color</th>
<th>L1 (µm)</th>
<th>L2 (µm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Red</td>
<td>11.06</td>
<td>9.51</td>
</tr>
<tr>
<td>Green</td>
<td>11.49</td>
<td>9.07</td>
</tr>
<tr>
<td>Blue</td>
<td>12.37</td>
<td>8.63</td>
</tr>
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cannot be differentiated in the NRR image. Figure 11 shows the geologic map of Kilian (1964) produced by boots on the ground mapping (before GPS and satellite imagery).

In general, there is good correlation but there is an obvious difference in the area south of fault b3 in Kilian’s map which shows the Cadiz Formation, while Figure 10 clearly indicates it is Wood Canyon Formation (WCF) or Zabriskie quartzite (ZQ). The geologic map of Kenney (2011) in Figure 12 more closely matches the NRR map and correctly shows the WCF/ZQ at this location but it may have benefitted from satellite imagery. The geologic map of Foster (2011) similarly shows the WCF/ZQ south of the fault but also shows additional faults not shown in the Kinney (2011) map, in an area of complex structure. The NRR image also easily discriminates the individual limestone/dolomite and shale/sandstone beds in the Cadiz Formation. The Latham shale is not well represented in the NRR image because it is easily eroded and commonly covered by talus from the overlying Chambless Limestone. While the geologic map is an idealized representation

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**Figure 11.** Geology of the southernmost Marble Mountains, from Kilian (1964). DCbk = Bonanza King Formation, Cc = Cadiz Formation, Cch = Chambless Limestone, Cpm = Wood Canyon Formation and Zabriskie Quartzite (formerly Prospect Mountain Quartzite), PCgr = Precambrian granite, TvB = Tertiary volcanic/basalt.

**Figure 12.** Geology of the southernmost Marble Mountains, from Kenney (2011). Ar = Archean basement, Wc = Wood Canyon Formation, Za = Zabriskie Quartzite, La = Latham Shale, Ch = Chambless Limestone, Ca = Cadiz Formation, BKi = Bonanza King Formation, Tv = Tertiary volcanics.
of the bedrock based on limited visual observations, remote sensing images display only the material exposed at the surface. Two alluvial fans map as forest green, which otherwise is not represented by actual outcrops in the NRR image. A subsequent ground truth survey showed that they consisted of desert pavement (Figure 13) comprised of a mixture of 7–30 cm clasts of the Chambless Limestone (blue in NRR map) and varnished Wood Canyon Formation or Zabriskie quartzite (light green in NRR map), the combination of which produces forest green on the 2-m resolution of the image. Figure 14 shows an aerial photo of the area and the tilted resistant gray limestone of the Chambless Formation can easily be discerned along with the tan shales and sandstones of the Cadiz Formation and reddish brown varnished Wood Canyon/Zabriskie quartzite.

Figure 15 presents the NRR image of the south-central Marble Mountains while Figure 16 is the geologic map of the area from Kilian (1964). The south-central Marble Mountains are less accessible than the area to the south of Route 66, so remote sensing has the potential for providing more detail and better accuracy than could
be covered on the ground. There appear to be significant areas where the Wood Canyon/Zabriskie Quartzite (light green in Figure 15) has been incorrectly mapped by Kilian (1964). In particular, the area southeast of the Iron Hat mine does not appear to show much WCF/ZQ in the NRR image, however, Kilian (1964) has mapped a large area of the unit there. In addition, at the extreme southernmost end of the image, there appears to be a partial sedimentary sequence in contact with the Precambrian granite but Kilian (1964) shows a direct contact between Tertiary volcanics and Precambrian granite. In contrast to the Precambrian granite, the Cretaceous quartz monzonite in the vicinity of the Iron Hat mine appears white to light purple in Figure 15. The Iron Hat mine (Figure 16) is a contact metamorphic deposit associated with the quartz monzonite and contains magnetite and hematite in addition to calcisilicate minerals (Lamey, 1948). The Tertiary volcanics show subtle color variations that may represent compositional differences that were not mapped by Kilian (1964).

Conclusions
The NRR maps of the southern Marble Mountains faithfully represented the geology previously mapped in the field but could not differentiate between very similar lithologies such as the Wood Canyon Formation and Zabriskie Quartzite and between carbonates in the Chambless Limestone, Cadiz, and Bonanza King Formations. It was able to detect several errors in the earliest geologic map of Kilian (1964). The ~2-m ground sampling distance of the remote sensing images allows lithologies to be mapped at much higher resolution than could be done practically in the field. The wavelengths used to create the NRR maps in Figure 10 and 15 did not take into account the laboratory reflectance measurements, which potentially could be used to fine tune the map by selecting wavelengths more unique to one lithology than another. Additional wavelengths could be chosen so that a series of NRR images could be prepared from the same data set to bring out as much differentiation as possible. This is being considered in ongoing work. The use of atmospheric compensated spectra could also improve the selection of wavelengths considered as background, as could examining the actual uncompensated spectra to understand the relative intensity of L1 wavelengths with respect to the corresponding L2 wavelengths.

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References


Remote sensing and mapping Miocene paleovalleys of the Marble, Bristol, and Old Dad mountains in the Trilobite and Bristol Mountain wildernesses, California

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ABSTRACT—Wilderness areas in the Mojave Desert, California, are remote and rugged terrain, but they contain important geology for understanding faults of the eastern California shear zone (ECSZ), and remote sensing offers techniques that can optimize mapping. The Bristol–Granite Mountain fault zone (BGMFZ) is the easternmost fault of the ECSZ with the Marble, Bristol, and Old Dad mountains on either side of the fault, as are the Trilobite and Bristol Mountain Wildernesses. In the northern Marble Mountains, a west-trending Miocene paleovalley has been proposed to have a correlative in the Old Dad Mountains and provides a constraint for right-lateral separation across the BGMFZ; however, this correlation is based on the premise that there was a unique paleovalley with a well-defined geometry. In the northern Marble Mountains, a paleovalley was mapped by the distribution of (1) thickness and facies within the Lost Marble gravel (LMg) and 18.8 Ma Peach Spring Tuff (PST), and (2) adjacent highlands where the PST was deposited on basalt and dacite lava flows. Whether this paleovalley is unique, or there are other paleovalleys farther south in the Marble Mountains, requires mapping of the entire 5 by 28 km area of Miocene volcanic rocks. In the south Bristol and Old Dad mountains, there is a similar 12 by 22 km area of Miocene basalt and dacite with deposits of PST and local sedimentary rocks, including the proposed offset Lost Marble paleovalley, but the entire range needs to be mapped to establish a unique correlate. The mountains are in the Mojave Trails National Monument, and the Trilobite and Bristol Mountains wilderness areas, so access is limited. Remote-sensing data, including aerial photography and hyperspectral images, are important for identification and characterization of rocks. Airborne hyperspectral Mako data can distinguish the distinctive spectral characteristics of the PST as well as several more map units identified by detailed field mapping in the Bristol Mountains. Reconnaissance maps derived from high spatial resolution Mako data can guide the detailed mapping needed to identify paleovalley or paleohighland deposits and can be used to optimize field time.

Introduction

The Trilobite and Bristol Mountains wildernesses include the Marble Mountains and Bristol and Old Dad mountains, respectively, and these mountains are remote and rugged, but they include rocks that have been proposed to constrain the amount of separation on the Bristol–Granite Mountain fault zone (BGMFZ). The BGMFZ is between the Marble Mountains (5 by 28 km) and Old Dad Mountains (12 by 22 km), for which there is no access other than backpacking. Throughout these mountains are scattered exposures of the Peach Spring Tuff, an 18.8 Ma widespread ignimbrite that filled topographic lowlands with <35 m thick deposits and mantled highlands with <10 m thick deposits. In the total extent of the PST, four general depositional environments have been identified: (1) local parts of the ignimbrite were deposited in a paleovalley, typically in medial to distal alluvial fan or lacustrine-playa environments but some were upper fan environments, (2) numerous PST deposits were isolated by erosion and valley margins cannot be identified, (3) paleovalleys where valley margins can be identified, and (4) areas with onlap deposits on paleohighlands (Buesch, 1991; Buesch, 2020). Where paleovalleys are well defined, the PST and substrate sediment (and the facies within these deposits) form the valley, and adjacent paleohighlands have PST deposited on lava flows, plutonic, or metamorphic rocks, and locally <1 m thick colluvium. Only by mapping these three rock units can the paleovalley be documented and the linear trend of the paleovalley axis and/or margin be determined. Lease and others (2009) identified the Lost Marble paleovalley in the northern Marble Mountains and proposed the correlative part of the paleovalley is in...
the Old Dad Mountain, and thereby determined ~17 km of right-slip separation on the BGMFZ (with an update in Miller and others, 2017) (Figure 1). The challenge for accepting this separation is to show that the Lost Marble paleovalley is unique in these mountain ranges, and this can only be proven by mapping the PST and subjacent rocks in these ranges. Using remote sensing, especially high-resolution hyperspectral imagery that can spectrally identify specific rock types and their spatial distribution along with LiDAR, enables assembling of reconnaissance maps that can help optimize field time.

Lease and others (2009) describe the Lost Marble paleovalley in the northern Marble Mountains as containing up to 30 m of the 18.8 Ma Peach Spring Tuff (PST; Ar-Ar age from Ferguson and others, 2013) and as much as 20 m of the underlying Lost Marble gravel (LMg). The 20.2 Ma Brown Butte dacite and overlying Castle basalt (Glazner and Bartley, 1990) form the base of the paleovalley and nearby highlands across which the PST was deposited.

Identifications of the PST and especially the LMg are critical for defining (and determining uniqueness of) the Lost Marble paleovalley in the northern Marble Mountains and the proposed separation to the Old Dad Mountains. Regionally, the PST is an important formation for stratigraphic, structural, and geochronologic control because it is a widespread ignimbrite exposed in many mountain ranges from the Barstow area (CA) to the western Colorado Plateau (AZ). There are examples of Miocene age paleovalleys containing the PST near...
Kingman (Arizona) and Kane Wash, southern Newberry Mountain (California) (Buesch and Valentine, 1986; Buesch, 1991). The PST was deposited from a large-volume pyroclastic flow, so the tuff formed in valleys and lapped onto or across highlands. The processes of deposition, compaction (welding), whether it remained vitric (glassy) or crystallized, and vapor-phase activity resulted in several facies that can be mapped. These facies can vary within valley-filling or highland-mantling deposits; therefore, they can provide important constraints on the depositional environment.

The LMg is poorly exposed, consists of pebble to cobble clasts (mostly dacite to andesite) in a sand-sized matrix, and varies in thickness from 0-20 m. Lease and others (2009) interpreted the LMg as a coarse-grained fluvial deposit, and the well sorted, textural maturity, and absence of any angular material suggests that the LMg was transported from distal sources; however, descriptions did not distinguish between axial fluvial deposits or local tributary deposits.

In many areas of the Marble, Bristol, and Old Dad mountains, the PST and local pre-PST sedimentary rocks are deposited on basaltic and dacitic lava flows and pyroclastic deposits, that can be locally highly variable in texture, composition, and pre-PST structural deformation, so detailed mapping can define the locations (and possible uniqueness) of paleovalleys and adjacent highlands to constrain separation across the BG MFZ. These mountain ranges are in the Mojave Trails National Monument, and the Trilobite and Bristol Mountains wilderness areas, so access is limited to long hikes and backpacking trips. Remote sensing, including aerial photographs and hyperspectral imaging (some with ground resolution of 1-2 m), are important for identification and characterization of outcrops in advance of field work. Reconnaissance maps derived from Mako data highlighting rock compositions can facilitate the detailed mapping needed to identify paleovalley or paleohighland deposits.

Geology of the northern Marble Mountains area

The northern Marble Mountains (Figure 2) includes exposures of Miocene age (1) Brown Butte dacite domes, lava flows, tuffs, and sedimentary sequences, (2) Castle basalt lava flows and cinder-scoria cones, (3) locally the Lost Marble gravel (LMg), (4) Peach Spring Tuff (PST), and (5) these rocks are overlain by epiclastic sedimentary rocks (Glazner and Bartley, 1990; Lease and others, 2009).

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Figure 2. Digital National Agriculture Imagery Program (NAIP) image of the northern Marble Mountains. Mesa labeled A is capped by the Peach Spring Tuff with the underlying Lost Marble gravel and forms paleovalley fill. Deposits labeled B have Peach Spring Tuff deposited on older volcanic rocks (no Lost Marble gravel) and forms paleohighland deposits.
One of the largest exposures of the PST forms a mesa on the ridge crest of the range, and this includes the thickest deposits of the LMg identified by Lease and others (2009) (labeled A in Figure 2 and pictured in Figure 3a). At this part of the mesa, the PST is 15–33 m thick, has a vitric and nonwelded base that grades upward into a vitric and moderately welded (v, mw) and then crystallized cliff-forming ignimbrite (site 3520). (C) Pebble sandstone in the LMg with all clasts in place (with the bounds of the book, pencils, and pens; site 3557).

Photographic imagery in the northern Marble Mountains

Aerial photography is important for mapping the distribution, thickness, and facies of rocks. Digital National Agriculture Imagery Program (NAIP) images have 1 m (pixel) ground resolution, and typically have three visible light bands (Red, Green, and Blue), although recent images include a fourth near infrared band (NIR). Using the R-G-B bands (NAIP bands 1, 2, 3, respectively, in RGB image channels) results in images that are very similar to color photographs (Figure 4a). Adding the near infrared band with visible band combinations, for example NIR-G-B bands (NAIP bands 4, 2, 3, respectively) as RGB image channels results in false color images that typically accentuate differences in rocks (Figure 4b).

Hyperspectral imagery

Many spaceborne and airborne sensors include geologically important spectral bands such as Visible (V), Near Infrared (NIR), Short-wavelength Infrared (SWIR), and Long-wavelength Infrared (LWIR) [also frequently termed as Thermal Infrared (TIR)] that are applicable to identification of rocks. Two examples illustrate the spectral characteristics of rocks.

One example (Figure 5) is from a small field site in a welded ignimbrite using a handheld ASD FieldSpec® Pro FR spectrometer† (Analytical Spectral Devices, 2002) that has spectral range and resolution characteristics equivalent to satellite and airborne instruments flown by NASA-Jet Propulsion Laboratory in Pasadena, California.
Figure 4. NAIP images of the PST, LMg, and older volcanic rocks at the mesa labeled as A in Figure 2. (A) Natural color RGB (NAIP bands 1, 2, 3). (B) VNIR False color with R as NAIP NIR band 4; G as NAIP band 2; B as NAIP band 3.

Figure 5. Photographs and hyperspectral profiles of absolute reflectance of sampling locations in the Topopah Spring Tuff (Tpt) on the west flank of Yucca Mountain. (A) and (B) are from the upper lithophysal zone (Tptpl). (C) and (D) are from the middle nonlithophysal zone (Tptpmn). Photographs are about 20 cm across. Standard deviations of the hyperspectral data are plotted, and wavelengths affected by absorption by atmospheric water are removed.
(Buesch and others, 2008). This instrument can record a complete 350 - 2500 nm spectrum (VNIR-SWIR) of a material at a sampling interval of 2 nm and a spectral resolution of 10 nm. As with the PST, the Topopah Spring Tuff (Tpt) is a nonwelded to densely compacted (welded) and vitric to crystallized ignimbrite. Both the PST and Tpt have lithophysal and nonlithophysal zones (lithophysae are gas cavities and associated crystallized parts of the rock), and the lithophysal and nonlithophysal zones in the Tpt were sampled for reflectance spectra. The ASD spectrometer does not result in an image, rather it collects the spectra from an area. At each of six selected locations, spectra were collected from five 1-cm-diameter sampling sites (area determined by the distance from sensor to object) within an area less than 30 by 30 cm (Figure 5). Because data were collected in natural light, atmospheric water resulted in absorption bands about 1350 to 1450 nm and 1780 to 1920 nm, so these wavelengths were not used in the reflectance profiles. The conclusion of this small study is that the lithophysal and nonlithophysal rocks have different spectral characteristics, especially at wavelengths greater than 800 nm.

A second example is from the Mako†# airborne LWIR/TIR hyperspectral sensor with spectral resolution of 128 bands over a wavelength range of the 7.8-13.4 µm acquired in August of 2013 at a spatial ground resolution of ~2.7 m (Hall and others, 2011). The RG swath (Figure 1) of Mako data covers 780 km across the south Bristol Mountains, Old Dad Mountains, and partial coverage of the northwestern Marble Mountains (Figure 1; Harvey and others, 2016). Data processing can extract high likelihood PST outcrops and lag deposits, highlighted in yellow and pink, respectively, on the south Bristol Mountains portion of the data. The subsequent OR swath of Mako data in the southern Marble Mountains shown in Figure 1 has been analyzed (Adams and others, 2021), but remains to be processed with techniques used in the Bristol Mountains. Based on compositionally dependent spectral characteristics (Figure 6), false color images can be derived to highlight specific compositional differences (Figure 7). The Mako images distinguish (1) a wide variety of volcanic, plutonic, and metamorphic rocks (Figure 7 and 8), (2) variations in the PST including vitric versus crystallized and vapor-phase crystallized or mineralized rocks (Figure 7 and 8), (3) variations in sedimentary rocks that can be attributed to compositional variation of clasts in the deposits and their respective provenance (Figure 9), and (4) separations along fault traces of conglomerate and strath terraces from the provenance of the clasts (Figure 9). The high spatial resolution of Mako can capture the presence of bedded strata underlying the PST despite the relatively narrow map view extent (Figure 8). These strata would not be distinguished from the compositionally similar underlying units in images derived from lower spatial and spectral resolution satellite or airborne sensors. Thus, Mako (and similar hyperspectral

![Figure 6. Spectral characteristics of six rock types from the sBM extracted from Mako data. PST1, PST3, PST3+carbonate, LS, and perlite are from locations on Figure 7. TQ cg and basaltic andesite are from south of the image. Each individual spectrum is an average of 5x5 pixels. Data are atmospherically corrected to apparent ground leaving radiation from which the emissivity information has been extracted.](image-url)
Figure 7. Oblique view of PST terrains in the south Bristol Mountains (sBM), false color 3 band minimum noise fraction (MNF) image. PST1 is a coherent ridge cap, PST2 is a sparse residual lag deposit (interpreted as disaggregation of formerly coherent welded tuff sheet), and PST3 is a coherent mesa cap rock. PST1 and 2 overly poorly indurated gypsum bearing sands and silts (s) overlying local Miocene volcanic rocks. PST3 overlies a generally coarsening upward sequence of gypsum bearing sandstones to conglomerate*, overlying and interbedded with local volcanic strata. PST facies are offset ~7 km across the right lateral SBMFZ and out of the figure. Plutonic rock (p), silicified rhyolitic lava (sFBR), perlitic rhyolite lava (pFBR), flow banded rhyolite (FBR), lacustrine or palustrine limestone (ls), volcanic ash and tuff (v). * Thinner and with less mature clast textures than the Lost Marble gravel (LMg).

Figure 8. Oblique view to SW of PST terrain in Old Dad Mountains (ODM). MNF image, colors do not correlate with Figure 7. Labels as in Figure 7. Rocks underlying the PST in the foreground are dominantly silicic dacites and locally derived rhyolites, no detailed geologic map exists. Arrows highlight the image capture of bedded strata underlying the PST. Bristol–Granite Mountain fault zone (BGMFZ)
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Conclusions
Establishing a well-constrained paleovalley (such as the proposed Lost Marble paleovalley of Lease and others, 2009) would provide a unique and linear geologic feature separated along a fault (such as the Bristol–Granite Mountain fault zone). Identification of rocks that formed in the paleovalley compared to those that formed the adjacent paleohighlands can constrain the margins of the paleovalley. Field work is essential for the identification of many of the crucial distinguishing stratigraphic features;
but remote sensing and especially use of high-resolution hyperspectral imagery and LiDAR provides invaluable reconnaissance by supplying preliminary data on the identification and spatial distribution of rock units based on spectral mapping. The combination of boots, hammers, LiDAR, and hyperspectral imagery offers the best combination for identifying and resolving whether there is a unique paleovalley offset along the Bristol–Granite Mountains fault zone.

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Eolian–fluvial interactions at the Kelso Dunes, Mojave National Preserve

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Abstract—Ephemeral washes blocked by dunes play a major role in modifying local hydrology and geomorphology. In the Mojave Desert, ephemeral washes that drain sediment from the Providence and Granite mountains were intercepted by the eastern margin of the Kelso Dunes. Temporary blockages of these washes by the dunes resulted in the deposition of about 10 meters of alluvium. Multiple exposures were described along the incised channel of Winston Wash and ages of sandy eolian and alluvial facies were determined by optically stimulated luminescence geochronology. The exposures revealed sequences composed of dune sand, sandy to gravelly alluvium, thinly bedded silt and clay beds, and alluvial fan gravels. Dune blockage and aggradation of alluvial fan channels occurred ca. 14–9 ka, consistent with the timing of regional alluvial fan sedimentation documented during the Pleistocene–Holocene transition. Incision of the dunes and aggraded sediments occurred during the mid-Holocene, correlated to a phase of alluvial fan activity 6–3 ka. Though temporary, dune blockages modify local base level of streams and can increase aggradation rates. They exemplify how eolian–fluvial interactions have influenced the geomorphology of the Mojave Desert.

Introduction

In deserts, active eolian dunes can interact with and block major axial drainages or smaller alluvial fan channels, resulting in significant sediment aggradation (Al-Masrahy and Mountney, 2015; Lancaster, 1993). This can be observed on the eastern margin of the Kelso dune field in Mojave National Preserve, southern California, where interbedded eolian and fluvial strata are exposed along Winston Wash cuts in the dune field (Figure 1). Lancaster (1993) was the first to document eolian–fluvial interactions in the eastern Kelso Dunes and speculated that dune blocking of Winston Wash occurred in the Holocene. Many researchers, including Lancaster (1993), suggested that the source of Kelso dune sand was from the Mojave River, a hypothesis later confirmed by Muhs et al. (2017). The increase in sediment supply that ultimately built the Kelso Dunes is related to the incision of Afton Canyon and the delivery of sediment from the Mojave River to the Lake Mojave basin ca. 25 ka (Reheis and Redwine, 2008). The sand was transported by wind along a sand transport corridor from the southern margin of Lake Mojave to the present location of the Kelso Dunes where they migrated up the alluvial fans of the Granite and Providence mountains (Lancaster and Tchakerian, 2003). Muhs et al. (2017) further documented that local alluvial fan sand sources contributed to the eastern parts of the Kelso dune field.

This paper focuses on the evidence for dune-blocked drainages along exposures of Winston Wash at the transition from the alluvial fans into the Kelso dune field.

Results

We described four sedimentary facies associated with eolian–fluvial interactions, similar to those associated with dune blockages in the Skeleton Coast Erg of Namibia (Svendsen et al., 2003). Optically stimulated luminescence (OSL) geochronology of sandy eolian and alluvial facies provided age control.

We described three stratigraphic sections along Winston Wash from Win-1 to Win-3, with Win-1 located close to the boundary between alluvial fans and sand dunes, and Win-3 located downstream and within the dunes (Figure 1a). The strata exposed along the wash contain multiple fining-up sequences of sand, silt, and clay with occasional gravel units. The facies, based on Svendsen et al. (2003), are as follows:

- Facies 1 consists of fine to medium grained, well-sorted, structureless or laminated sand that is interpreted as eolian in origin.
- Facies 2 consists of fining-up sequences with thin beds of fine to coarse laminated sand at the base, followed by silt and clay beds. The clay beds contain up to 60% clay-sized sediment, and some are capped with mud cracks. At the outcrop scale, the fining-up sequences fill channels cut into facies 1 dune sand and are roughly 50 m wide. Individual fining-up sequences are typically about 20 cm thick but can be up to 50 cm thick.
- Facies 3 consists of sand to gravel beds, up to 1 m thick, representing fluvial transport. Sand beds may contain mud rip-up clasts or slump blocks composed of stratified...
dune sand. These beds can be found associated with eolian sand of facies 1 or channel fill of facies 2.

Facies 4 is composed of gravel beds with normal or reverse grading deposited by stream flow or debris flow on alluvial fans. These gravel beds are mostly found at the base of or capping channel fill deposits of facies 2 and/or 3.

An example of some facies relationships at site Win-1 is depicted in figure 1d. This site is dominated by facies 1 dune sand and facies 2 fining-up sequences. Facies 1 is typically overlain by facies 2. Facies 3 fluvial sediment and facies 4 gravels are rare but associated with any facies. At Winston Wash, facies alternate between 1 and 2 at the transition from alluvial fans to sand dunes. As the wash progresses further into the dunes the stratigraphic sections (Win-3; see Sweeney et al., 2020) become dominated by facies 1.

OSL ages from Win-1 (Fig. 1d) are from facies 1 basal eolian sand (ca. 14 ka) and facies 4 sandy alluvial fan sand (ca. 10 ka) capping a fining-up sequence. OSL ages from other dune blockage sites in the Mojave Desert cluster around the Pleistocene–Holocene transition (Sweeney et al., 2020). The 10 ka alluvial fan unit at Win-1 is overlain by an additional 6 m of facies 1 and 2, and capped by stabilized dunes dated at ca. 8 ka (Lancaster and Tchakerian, 2003).

Discussion

Stratigraphic facies relationships and age control at Win-1 suggest that sand dunes had migrated up alluvial fans of the Providence Mountains by ca. 14 ka and blocked alluvial fan drainages, resulting in periodic aggradation of alluvium alternating with incursions of eolian sand until eolian sand deposition dominated at 8 ka. We interpret each fining up sedimentary sequence in the aggraded alluvium as representing individual waning flood flows with the coarsest sediment deposited first, and with slackwater or ponding resulting in clay deposition last. The high concentration of clay (up to 60%), high salinity comparable to other dry lake beds (Sweeney et al., 2020), and mud cracks associated with some clay layers, suggest

Figure 1. A. Google Earth image of the field area where Winston Wash dune blockage strata are exposed. Drive 5.1 miles (8.2 km) south of Kelso, park near the “Dip” sign and hike ~2.25 km west of Kelbaker Road. Locations of other sites showcasing dune dam strata documented in Sweeney et al. (2020) are also shown. B. Map showing location of field site within the Mojave National Preserve. C. Photo facing east of Facies 2 channel fill exposure and Facies 4 alluvial fan gravels exposed at Win-1. Providence Mts. in the background. D. Composite stratigraphic column of the Win-1 site depicting the different sedimentary facies and luminescence ages.
that temporary bodies of water may have existed prior to full evaporation. The facies 4 alluvial fan gravel that interrupts the sequence at Win-1 is likely related to a large pulse of alluvial fan sedimentation at the Pleistocene–Holocene transition (McDonald et al., 2003) representing higher energy flood flows, compared to the flows that formed facies 2 and 3.

Site Win-2, downstream from Win-1, is dominated by facies 2 fining-up sequences, with nearly every fining-up sequence capped by clay. The strata at Win-2 likely represent the progressive infilling of one alluvial fan channel by successive flood events. Site Win-3, the farthest downstream into the dune field, is dominated by facies 1 dune sand with occasional interbeds of facies 2, 3, and 4. Win-3 likely represents ephemeral streams that were rarely able to penetrate into or through the dune field.

Some facies provide evidence of the erosion of sand dunes by flood flows. An exposure just downstream of Win-1 contains rotated 20 cm blocks of dune sand encased in poorly sorted alluvial sand (Facies 3) that may have resulted from fluvial erosion and undercutting of dunes followed by the collapse of blocks of semi-cohesive (moist?) sand into the channel (c.f. Svendsen et al., 2003). In addition, some facies 3 exposures are composed of well-sorted sand likely derived from the reworking of eolian sand but containing scattered granule to pebble layers or mud rip-up clasts.

The timing of dune blockage and aggradation correlates to a significant episode of alluvial fan activity (Qf5, McDonald et al., 2003; Qya4, Miller et al., 2010). Dune blockages of streams resulted in the aggradation of >10 m of alluvial and eolian sediment at Winston Wash at a rate of 1 to 3 m per thousand years. This aggradation increased local stream base levels. The gradient of Winston Wash during the aggradation phase behind the dune dam, determined by tracing correlative beds over several hundred meters of exposure, was 16 m/km, compared to 26 m/km of the present-day channel. This reduction in gradient occurred as channels filled with sediment and led to the ultimate breaching and incision of the dune dam when accommodation space for sediment was reduced. Our luminescence chronology and stratigraphic relationships suggest that the main alluvial aggradation phase occurred ca. 14–9 ka, followed by eolian activity in the early to mid-Holocene (Lancaster et al., 2003) that resulted in the migration of dunes over dune-dammed channel fill and alluvial fan deposits.

Breaching of the dune dam and the incision of the Winston Wash channel likely occurred with the next major phase of alluvial fan activity, ca. 6–3 ka. The timing for the breach is interpreted from the 8 ka dune age overlying the Win-1 deposits (Lancaster and Tchakerian, 2003) and the presence of Qf6 (6–3 ka; McDonald et al., 2003) alluvial fan deposits adjacent to the active Winston Wash channel where it cuts through the dunes, inset into the dune blockage deposits. This timing is also corroborated by evidence found in deposits along Kelso Wash downstream (Sweeney et al., 2020).

Al-Masrahy and Mountney (2015) classified different types of eolian–fluvial interactions, and several examples of modern and late Pleistocene interactions can be observed in the Mojave Desert via strata or modern geomorphic observations. The present-day Cottonwood and Winston washes flow through the entire Kelso dune field unimpeded in part because portions of the dune field are stabilized with vegetation, thus leaving the ephemeral channels open. Devils Playground Wash, on the southern margin of the Kelso dune field, flows north from the Granite Mountains and appears to be blocked and diverted to the west around the edge of the dune field. During the Pleistocene–Holocene transition, however, alluvial fan channels may have entered the dune field via interdune corridors or ponded against larger dune forms. According to Al-Masrahy and Mountney (2015), if water levels in ponded areas rose to the elevation of swales or saddles in the dune topography, water could spill over into adjacent interdunes. Contemporaneous flood flows during the Pleistocene–Holocene transition likely occurred across the Providence Mountains bajada and entered the dune field at multiple locations.

Langford (1989) documented other recent eolian–fluvial interactions at the Mojave River delta area that included smaller-scale dunes that blocked channels, ponding, and subsequent erosion related to flooding on the Mojave River. Other dune-blocking events have been documented along Kelso Wash, related to the increase in sediment supply from the Mojave River and pulse of eolian sand blown through the Devils Playground area from the Mojave River ca. 20 ka (Sweeney et al., 2020). Dune blockages on larger drainages such as Kelso Wash resulted in enhanced aggradation upstream of the blockage while promoting relative landscape stability and soil development downstream of the blockage that likely lasted several thousand years. OSL chronology (Sweeney et al., 2020) suggest that the blockage of Kelso Wash occurred at the same time as the blockage of Winston Wash suggesting that the migration of dunes that built the Kelso dune field, coinciding with phases of alluvial fan activity, had a profound impact on the local geomorphology via the modification of base level. Eolian–fluvial interactions are abundant in the eastern Mojave Desert and have played an important role in the landscape evolution of this region.

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Disjunct chaparral relicts in the New York Mountains of Mojave National Preserve: a preliminary survey

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Abstract—Disjunct populations of relict chaparral plant species occur in isolated montane sites across the Mojave Desert, including those found in the New York Mountains of Mojave National Preserve. In this study, the plant community composition of sites where several of these species occur were censused in both the New York Mountains and within the California Floristic Province in the Transverse Ranges. The relative abundances of plant species found were used to (1) compare differences in the diversity and evenness of sites between the ranges; (2) determine if the disjunct species mean relative abundances appear to differ between the New York Mountains and Transverse Ranges; (3) characterize the assemblages of sites within the ranges using a clustering analysis; and (4) compare the composition of sites using a multivariate ordination analysis. Results indicated a non-significant difference in both the diversity and evenness between sites in the New York Mountains and Transverse Ranges, with evenness appearing to be an important factor related to diversity in the Transverse Ranges. There was a marginally non-significant difference in the mean relative abundances of disjunct taxa between the mountain ranges. Clustering analysis pointed towards two types of community assemblages associated with pine-oak woodland and chaparral shrub species found in these ranges. The ordination indicated that there was a significant difference in the composition of sites between ranges being driven by a few species associated with the woodland assemblage type. These preliminary results serve as the baseline for future analyses that aim to explain the distribution of the relict plant species in a broader study including additional mountain ranges across the Mojave Desert.

Introduction
The New York Mountains in Mojave National Preserve represent an ecological confluence where plant species characteristic of distant floristic regions co-occur (Jordan 2020). Among these are species native to the coastal chaparral habitats of the California Floristic Province (CFP), a region of especially high plant diversity (Raven & Axelrod 1978; Thorne et al. 2009; Kraft et al. 2010; Lancaster & Kay 2012; Burge et al. 2016; Baldwin et al. 2019). While the Mojave Desert does not fall within the CFP, the occurrence of CFP native species within the New York Mountains is an interesting exception. Disjunct populations of plant taxa associated with California’s coastal chaparral occur in just a few pinyon-juniper woodlands found within the Mojave Desert, with some of the most accessible observations recorded in the mountains of Mojave National Preserve. Although the structure and composition of a few of these woodland communities has been documented (Brown 1978; Prigge 1979; Trombulak and Cody 1980; Cardiff and Remsen 1981; Thorne et al. 1981; Andre 2006), the populations of disjunct taxa have not yet been compared among desert mountain ranges, or to those found in the more widespread coastal chaparral communities. These populations warrant conservation attention because of threats posed by climate change, wildfire, and invasive species that could alter the available habitat (Engel and Abella 2011; Abella 2012; Elsen and Tingley 2015).
Evidence from pack rat (Neotoma spp.) middens has provided clues to suggest how the flora of what is now the California Desert has changed since the late Pleistocene, and it appears that a wetter climate allowed CFP species to be distributed across a broader landscape than they presently occupy (Wells and Berger 1967; Van Devender 1977; Spaulding 1991; Smith et al. 2000; Holmgren et al. 2014; Balmaki and Wigand 2019). Presumably, many conifer woodlands once more widespread became stranded on mountaintops via habitat tracking as the Mojave Desert dominated part of this landscape within the last 10,000 years (Smith et al. 2000; Enzel et al. 2003; Cole et al. 2008). Likewise, it appears the chaparral species that occur in the New York Mountains may have followed a similar pattern that is permitted in part by a unique climate driven by increased elevation and bimodal annual precipitation influenced by the North American monsoon (Tubbs 1972; Adams and Comrie 1997; Hereford et al. 2006). This might be the reason that, while many high elevation mountain ranges within the Mojave Desert currently harbor conifer woodlands, chaparral plant species seem to occur mostly in ranges in the eastern part of the region where monsoonal precipitation is more prevalent.
This project seeks to examine the composition of chaparral and woodland plant communities in the New York Mountains and within the CFP in the San Gabriel and San Bernardino Mountains of the Transverse Ranges to explore patterns in the distribution of these chaparral species and their associated flora. To explain the distribution of chaparral populations within this landscape, I aim to address these questions: (1) Are there differences in the plant community diversity and evenness between sites in the New York Mountains and Transverse Ranges? (2) Do the disjunct taxa in the New York Mountains communities have abundances similar to their abundances within more widespread coastal communities in the Transverse Ranges? (3) Are populations of these disjunct taxa associated with characteristic assemblage types, and if so, with which dominant plant species are they associated? (4) How does the plant species composition of these communities compare among sites within the mountain ranges?

Answering these questions will elucidate how such assemblages are structured and support predictions about where the disjunct taxa are expected to occur within the Mojave Desert. The relationships between diversity and evenness within these communities can illustrate patterns related to vegetation structure where the chaparral populations occur. If there are differences in the relative abundances of the disjunct taxa between the New York Mountains and Transverse Ranges it could reflect lower dispersal, recruitment, and fecundity related to their isolation in the Mojave Desert. Assigning populations of the co-occurring plant species found in these ranges to specific assemblage types might reveal variation in the kinds of environments influencing distribution of the disjunct taxa. Likewise, comparing the composition of assemblages found within these ranges illustrates patterns of biogeographic change across the landscape where these populations occur.

Methods

In June and July 2020, sites for conducting belt transects (50 m long; 1 m wide) in the New York Mountains (n = 7 transects), and Transverse Ranges (n = 7 transects) were selected haphazardly based on the presence of disjunct taxa known from California Consortium Herbarium records (https://ucjeps.berkeley.edu/consortium/) and ease of accessibility (Fig. 1; Table 1; Table 2). Within the Transverse Ranges, sites were selected in both the San Bernardino (n = 3 transects) and San Gabriel Mountains (n = 4 transects). The relative abundances of all living plants found to have any canopy overlap across or above the 1 m belt were recorded along each transect. Species abundance curves produced following these surveys indicated that few additional plant species were encountered as the additive length of the transects surveyed approached 350 meters (7 transects) in the New York Mountains, and likewise in the Transverse Ranges.

The relative abundances of plant species recorded in transects were used to quantify the diversity and evenness of each community. Diversity was calculated using Shannon’s index $H'$, where $H' = -\sum (p_i \ln(p_i))$, and $p_i$ is the relative abundance of a given species found. Shannon’s index combines species richness and relative abundance to provide a measure of diversity that increases with the richness and similarity of relative abundances among
species. Evenness was calculated as $H'/H_{\text{max}}$ (where $H_{\text{max}} = \ln(r)$, and $r = \text{species richness}$), and provides a measure of how similar each of the relative abundances are; evenness is equal to 1 when all relative abundances are the same.

Differences in diversity and evenness between the New York Mountains and Transverse Range sites were both compared using two sample t-tests (Welch’s test assuming unequal variances). Both of these data sets met the assumptions of normality (Shapiro test $P > 0.05$). Species richness, evenness, and percent cover (calculated as the total cover observed in cm divided by the length of the transect, 5000 cm) were each tested as correlates with diversity in Pearson correlation coefficient analyses to evaluate if these factors might be related to diversity. Species richness and evenness were similarly analyzed for both ranges. The mean relative abundances of the disjunct taxa (Table 1) as well as Single-leaf Pinyon Pine (Pinus monophylla), which frequently occurred in sites, were then plotted between the New York Mountains and Transverse Ranges to illustrate how the distribution of each may differ between ranges. A paired t-test was used to compare the differences in these two sets of species abundances, with each of the nine species paired between ranges.

To evaluate which plant taxa tend to co-occur in these sites, relative abundances of the community dominants (Table 2) were used in a k-means clustering analysis, performed with the cluster and factoextra packages in base R (The R project for statistical computing, version 4.0.2). k-means clustering is a non-hierarchical clustering algorithm that here analyzes the likelihood of these dominant taxa occurring in similar types of community assemblages. This k-means analysis used the average silhouette method as an objective technique for identifying the appropriate level of clusters ($k$) to explore.

An ordination analysis was used to compare the composition of sites within each of the three ranges (New York, San Bernardino, and San Gabriel). Relative abundances of the community dominants (Table 2) were used to conduct Nonmetric Multidimensional Scaling (nMDS) and Permutational Multiple Analysis of Variance (perMANOVA), both performed with the package vegan in R. nMDS is an ordination method that can graphically illustrate which ranges have the most similar species composition within the sites surveyed. perMANOVA is a multivariate linear model that here takes the three mountain ranges as levels of the independent variable, and the dominant species relative abundances as multiple factors of the dependent variable. The results of perMANOVA show how much the species composition of sites across the three ranges varies, and which species occurrences are most strongly related to these differences.

A formal discussion of the statistical parameters used to execute k-means clustering and nMDS analyses is described by Quinn and Keough (2002).

### Results

The difference in diversity between the New York Mountains and Transverse Range sites was marginally non-significant ($t = 1.813$, df = 10, $P = 0.070$; Fig. 2), and there was not a significant difference in evenness ($t = -1.729$, df = 10, $P = 0.110$; Fig. 2). Diversity and species richness were positively correlated among sites in the New York Mountains ($r = 0.942$, df = 6, $P = 0.002$; Fig. 3), but not correlated among sites in the Transverse Ranges ($r = 0.598$, df = 6, $P = 0.156$; Fig. 3). Diversity and evenness were positively correlated in both the New York Mountain ranges as levels of the independent variable, and the dominant species relative abundances as multiple factors of the dependent variable. The results of perMANOVA show how much the species composition of sites across the three ranges varies, and which species occurrences are most strongly related to these differences.
Mountains (r = 0.942, df = 6, P = 0.002; Fig. 3), and the Transverse Ranges (r = 0.967, df = 6, P < 0.001; Fig. 3). Diversity and percent cover were positively correlated in the New York Mountains (r = 0.768, df = 6, P = 0.044; Fig. 3), but not correlated in the Transverse Ranges (r = -0.225, df = 6, P = 0.628; Fig. 3). Evenness and species richness were positively correlated in the New York Mountains (r = 0.777, df = 6, P = 0.040; Fig. 4), but not correlated in the Transverse Ranges (r = 0.374, df = 6, P = 0.408; Fig. 4). There was a marginally non-significant difference in the mean relative abundances of the nine species compared between New York Mountains and Transverse Ranges (t = 1.954, df = 8, P = 0.087; Fig. 5).

The average silhouette width indicated that k = 2 was the most appropriate level of k to explore, with clusters of size n = 5 and n = 16 species identified (Fig. 6). Cluster 1 was characterized by species associated with a pine-oak woodland assemblage type, whereas Cluster 2 was characterized by species associated with a chaparral shrub assemblage type. Although there were some species assigned to the chaparral cluster expected to be associated with pine-oak woodland (e.g. Juniperus osteosperma and Pinus edulis), species were generally assigned to the cluster most representative of the type of plant assemblages where they were found.

The nMDS plot showed marginal overlap in species composition between the New York and San Gabriel mountains, while the San Bernardino Mountains showed no overlap with either of the other two ranges (stress = 0.170; Fig. 7). There was a significant difference in the species composition among the ranges (F_{2,31}, P = 0.001). The perMANOVA coefficients indicated that Pinus monophylla, Quercus chrysolepis, and Garrya flavescens were strongly associated with differences in the composition of sites between ranges.

Discussion

The similarity in diversity and evenness between these mountain ranges might illustrate similarities in community structure and distribution that reflect idiosyncratic environmental factors occurring in different regions across the landscape. Significant correlations between several variables (species richness, evenness, percent cover) and diversity in the New York Mountains indicate that multiple factors in community structure are likely influencing patterns of diversity there. The single significant correlation between evenness and diversity in the Transverse Ranges suggests that diversity there is instead related to a relatively homogeneous community structure rather than to species richness or percent cover. Likewise, significant correlations between species richness and evenness were only found in the New York Mountains, suggesting that multiple variables are related to patterns of community structure there. While there was a marginally non-significant difference in the mean relative abundances of disjunct taxa between the two mountain ranges, there was a higher abundance of a few species (e.g. Garrya flavescens, Rhus aromatica, and Rhamnus ilicifolia) in the Transverse Ranges (Fig. 6). These differences might reflect variable patterns of fecundity and dispersal, or perhaps a response to competition in different environments.

Previous work in Mojave National Preserve that subjectively used species composition to describe highland plant communities generally agree on patterns of vegetation structure (Prigge 1979; Thorne et al. 1981; Cardiff and Remsen 1981; Andre 2006). Prigge (1979) and Thorne et al. (1981) similarly described three communities in the New York Mountains where chaparral species occur that were generally characterized as juniper-sagebrush scrub, pinyon-juniper woodland, and pinyon-oak woodland. By comparison, the k-means analysis that generated two clusters here supports the occurrence of pine-oak woodland and chaparral shrub assemblage types that include the disjunct taxa in both the New York Mountains and Transverse Ranges (Fig. 6). However, some species assigned to these clusters did not exclusively co-occur in such assemblage types. For example, a few of the disjunct taxa in the New York Mountains found in pine-oak woodland settings such as Rhamnus ilicifolia and Ceanothus pauciflorus were instead assigned to the chaparral cluster. Such results show that populations of these disjunct species might tend to favor less stressful mesic sites in their disjunct environment, while instead inhabiting more xeric sites in coastal mountain ranges where their distributions are more widespread. Differences in the assignment of disjunct taxa to these clusters suggests that their distribution likely
Figure 3. Correlations of diversity with species richness, evenness, and percent cover compared between sites in the New York Mountains and Transverse Ranges (n = 7 sites for each mountain range).
depends upon variable environmental factors across this landscape.

Altering the levels of $k$ somewhat more subjectively could expand these interpretations. For example, with $k = 3$, the third cluster includes *Garrya flavescens* and *Frangula californica*, which both were found in alluvial drainages that might be characterized as yet another assemblage type. The latter species was exclusively found in such sites in all three mountain ranges. Removing the conifer *Juniperus osteosperma* from the chaparral cluster requires setting the level of $k = 6$, and groups it with *Eriodictyon angustifolia* and *Rhus aromatica*. Sites where these three species were found to co-occur in the New York Mountains were somewhat structurally similar to a chaparral shrubland, and probably represent the juniper-sagebrush scrub described by Thorne et al. (1981). At this same level of $k$, *Quercus chrysolepis* is removed from the pine-oak woodland and independently assigned to a unique cluster; the k-means analysis only groups *G. flavescens* with *Q. chrysolepis* in a pine-oak assemblage at $k = 2$. Sampling more sites within mountain ranges across the Mojave landscape will be necessary to ultimately resolve these uncertainties and indicate if any additional clusters can be identified with greater confidence.

Differences in the species composition between these ranges might reflect variation related to environmental influence on vegetation structure across the landscape. Results of perMANOVA suggested that the distribution of *Pinus monophylla*, *Quercus chrysolepis*, and *Garrya flavescens* were driving the differences among the ranges (Fig. 7) consistent with patterns shown by k-means clustering (Fig. 6). These taxa were found in sites that are likely influenced by important environmental factors allowing assemblages to become dominated by woodland species. Perhaps species of the chaparral and woodland assemblage types are tracking similar microsites that tend to overlap within these mountain ranges. It is possible that the formation of a dense and elevated canopy structure impedes the ability of many shrubs to become established in microsites where dominant woodland species are present; diversity and percent cover were negatively associated across these mountain ranges which might be evidence of this phenomenon (S. Jordan, unpublished data). Variation in the topographic position of sites (e.g. ridges vs. drainages) likely exerts influence on the kind of assemblages that have formed (e.g. Pinder III et al. 1997). Elevation and aspect influence soil moisture availability (Parker 1982) and are additional environmental factors that are likely to underly the distribution of these species. Although dispersal patterns were not
explored here, the chance movement of seeds between sites, or more broadly among mountain ranges, might explain some of the variation in these distribution patterns as well.

Climatic influence on the distribution of chaparral and woodland plant communities across this landscape cannot be overlooked. The North American monsoon, while less frequent in California than further east, has afforded some parts of the eastern Mojave Desert a bimodal precipitation trend (Tubbs 1972; Hereford et al. 2006; S. Jordan, unpublished data). The timing of this late season rainfall may have a significant influence on vegetation patterns in mountain ranges within the eastern Mojave Desert, because it provides relief from the seasonal drought that is characteristic of the Mediterranean climate in Southern California. While precipitation generally increases with elevation regionally, it is likely that these eastern ranges will receive more annual rainfall compared to ranges at similar elevations further west in the Mojave Desert (S. Jordan, unpublished data).

Figure 6. k-means cluster plot with Principal Components Axes. Average silhouette width indicated that species are best assigned to two clusters; one appears to be characterized as a pine-oak woodland assemblage (Cluster 1) and the other a chaparral shrub assemblage (Cluster 2). Taxon codes are specified in Table 2.

Figure 7. Non-metric Multidimensional Scaling (nMDS) ordination of plant species composition among ranges. Proximity between the species represents similarities in their distribution. Taxon codes are specified in Table 2.
In fact, the chaparral and woodland species found in the New York Mountains (Figure 8) are completely absent at the same elevations in western Mojave ranges such as the Ord Mountains (personal observation). This additional precipitation along with variable topographic positions amenable to retaining moisture appear to provide refugia that support woodland and chaparral species within parts of these otherwise more arid eastern desert mountains. Considering that these plant communities are isolated relicts of the wetter Pleistocene times, they deserve study as impending climate stress alters the availability of suitable habitat.

Conclusion

Preliminary results here illustrate some of the ecological characteristics associated with the distribution of these disjunct chaparral taxa. The plant communities sampled appear to have similar diversity, evenness, and relative abundances of disjunct taxa perhaps related to similar environmental influence, however there are perhaps a few important environmental factors that do vary among ranges. Patterns of co-occurrence between species seem to exist, and there are roughly discernable assemblage types that can be identified among these ranges. The differences in species composition among ranges observed here likely reflect variation in biogeographic patterns that exist across the Mojave Desert. Censusing more ranges in the eastern Mojave Desert will generate additional data that is needed to better understand these patterns. Examining environmental variables that quantify topographic position and soil moisture characteristics will be a useful next step in identifying variation in important factors influencing species distribution across these ranges.

Regional biodiversity in the Mojave Desert continues to be threatened as species face extirpation or extinction due to habitat loss, especially those with limited distribution (Hunter et al. 2003; Guida et al. 2019). Investigating the occurrence of these isolated populations enables us to better understand the disjunct species distributions and variation in the composition of community assemblages where they occur. Such evidence provides further support for maintaining protection of plant populations within the Mojave National Preserve while emphasizing how plant diversity in the New York Mountains highlights the species richness of the preserve. Identifying factors influencing the distribution of these populations would allow further predictions to be made about how the range of disjunct taxa within the Mojave Desert will be expected to shift in the future as warming and aridity increase. Should subsequent models predict these taxa will contract their range as their fundamental niche becomes smaller, it would inform the management and conservation of isolated plant populations within Mojave National Preserve.

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Literature cited


Figure 8. New York Mountains near the mouth of Caruthers Canyon. J. Reynolds photograph.


Hyperspectral Thermal Emission Spectrometer (HyTES) images of basaltic and sedimentary deposits in the southwest Cima volcanic field, California

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ABSTRACT—The southwestern part of the Cima volcanic field in the Mojave National Preserve, California, contains many of the youngest basaltic cinder cones and lava flows in the field (Wilshire and others, 2002). In 2014 the Hyperspectral Thermal Emission Spectrometer (HyTES) collected a swath of data across this area. This summary describes the HyTES instrument, data, and images, and compares two standard images (L1 Brightness Temperature and L2 Land Surface Temperature) to the geologic map and aerial photographs. In aerial photographs (visible spectrum) and HyTES images (thermal infrared spectrum), there are very good correlations of features such as types of cones, pāhoehoe and `a`ā flows (smoother or rougher surfaces, respectively), channelized features such as levees and flow-induced arcuate ridges across the channels, some flows with rafted parts of the source cinder cone, and different types of sedimentary deposits. The comparison is visual and qualitative; however, it shows the potential for applying hyperspectral data in a quantitative characterization and identification of lava flows and cones, sedimentary deposits, and locally exposed bedrock gneiss.

Introduction
The southwestern part of the Cima volcanic field in the Mojave National Monument, California, contains many of the youngest basaltic cinder cones and lava flows in the field (Wilshire and others, 2002; Figure 1). Mapping by Wilshire was conducted in the field and aided by aerial photographic interpretations, geochemical analyses, and geochronology. Recently acquired aerial photographs are available through Google Earth†, and digital aerial images are available through the National Agriculture Imagery Program (NAIP). Aerial photographs in Google Earth are typical photographs (or images) based on three visible light bands (Red, Green, and Blue, R, G, and B, respectively), but digital versions of individual bands are not available and are effectively “flat” analog photographs. NAIP images have three visible light bands (R, G, and B) with recent images having an additional near infrared band (NIR). The Hyperspectral Thermal Emission Spectrometer (HyTES) is operated by the National Aeronautics and Space Administration (NASA) Jet Propulsion Laboratory (JPL). In 2014, the HyTES collected a swath of data across the southwestern part of the Cima volcanic field (Figure 1). This summary describes the HyTES instrument, data, and images, and compares two standard images to the geologic map and aerial photographs. The HyTES images depict spectral thermal data (L1 Brightness Temperature and L2 Land Surface Temperature) as a single layer (flat) red-green-blue image; however, they are based on the digital data. The comparison is visual and qualitative; however, it shows the potential for applying hyperspectral data in a quantitative characterization and identification of lava flows and cones, sedimentary deposits, and locally exposed bedrock gneiss.

HyTES instrument, data, and images
HyTES is an airborne pushbroom imaging spectrometer with 256 contiguous spectral channels between 7.5 and 12 micrometers (µm) in the thermal infrared part of the electromagnetic spectrum and 512 pixels in the cross-track dimension. Characteristic thermal infrared spectra are indicative of specific rock, alteration, and soil compositions that are typical in other multispectral thermal infrared images (Hook and others, 2005). HyTES data have been acquired since 2013 from multiple airborne platforms, with ground resolutions between 1 and 36 m depending on flight altitude. The Cima data have a 6.24 m ground resolution, and were acquired 2014-07-06 ~4:45 pm PDT, so east-sloping sides of cones and other areas of relief are shaded and cooler. HyTES data are typically analyzed using special imaging software, but they are initially visualized as two image files titled “L1 Brightness Temperature” (L1 BT RGB) and “L2 Land Surface Temperature” (LST). The L1 BT RGB image (Fig. 1B) has RGB Red band with channel 150 (10.1 µm), Green...
Figure 1. Geologic map of the southwest part of the Cima volcanic field (A), and HyTES L1 BT RGB image overlay on Google Earth aerial photograph (B). Undivided gneiss (Xu), slide block of interbedded quartzite, shale and dolomite or limestone (Tsq), vents (V), tephra and lava flows (QT), and sedimentary deposits (Q) are labeled as in Wilshire and others (2002). Selected contacts are drawn to denote the boundaries of different deposits. Black numbered dots on geologic map represent the age of the rock (Wilshire and others, 2002).
Band with channel 100 (9.2 µm), and Blue Band with channel 58 (8.5 µm) (Figure 1). The L2 LST image (Fig. 2) is scaled with dark blue <309 K to dark red >336 K. Cima images were subjected to the standard processing of all HyTES data, and no image- or rock-specific processing was carried out. A slight color shift from south to north apparent across the swath can be normalized with additional processing.

Comparison of geologic map, aerial photographs, and HyTES images

The geologic map of the southwest Cima field includes Proterozoic bedrock (gneiss) and Tertiary to Quaternary sedimentary units (Wilshire and others, 2002) (Figure 1) as well as the basaltic rocks that are the focus of this paper. The study area includes Early Proterozoic mafic and felsic gneiss (Xu) and Miocene sedimentary breccia (Tsq). The basaltic rocks are divided into numbered vents (V) and flows. The vents are composed of pyroclastic deposits (ash, scoria, and bombs) that typically form cinder or scoria cones, and the lava flows are labeled by their source vents (for example, QTf2-1 is the numbered flow 1 from vent 2). Most pyroclastic vents in the Cima field are steep-sided, high-relief cones with a small diameter crater relative to the diameter of the base of the cone, and these scoria cones resulted from magmatic Strombolian eruptions (V11 is a good example, Fig.1). Scoria cones can be the vents for lava flows that emanate from the vent crater, as satellite vents or fissures on the side or base of the cone, or from the vent of a breeched cone. Less described are low-relief, shallowly to moderately dipping cones with a large diameter crater compared to the diameter of the base of the cone, and these tuff cones, tuff rings, or maars resulted from phreatomagmatic eruptions. Vent 12 (Fig. 1) has an early formed, dark gray tuff ring, and three red-brown scoria cones. Sedimentary units are typically alluvial deposits and mapped based on grain sizes, geomorphic process, height above adjacent drainages, and soil development (symbols Q with numbers and in some cases letters).

Overall, there is good agreement in comparisons of mapped units to aerial photographs in Google Earth and also to the HyTES L1 BT RGB (Fig. 1) and L2 LST images. Two, small, mapped areas in the HyTES swath have distinctive photographic and HyTES image characteristics, one being of map unit Xu (gneiss), and the other of unit Tsq (a slide block composed of interbedded quartzite, shale and dolomite or limestone) (Wilshire and others, 2002). For the lava flows, the aerial photographs depict lighter and darker tonal variations consistent with pāhoehoe and ‘a‘ā flows and smoother or rougher surfaces (respectively). Many flows have channelized features such as levees and flow-induced arcuate ridges across the channels, and some flows have rafted parts of the source cinder cone. Many of these features can be identified in HyTES data and images. HyTES data
indicate compositional and/or textural differences in lava flow units mapped as QTf2-1, QTf4-2&3, QTf7-1&3&4, QTf12-1, and two not-identified flows QTfx and QTfz. Lava flow QTf6B-1 can be traced for ~5.5 km to V6B that is adjacent to the scoria cones V6A and V7. However, V6B is a low-relief, large-diameter crater, probably a maar, that was filled by lava and spilled to form the flow, and parts of the QTf6B-1 flow are partially overlain by alluvial deposits Q2b. HyTES images depict lava flow features such as an `a`ā flow with large arcuate ridges and rafted cinder cone fragments (QTf2-2-1), and a lava-flow channel with arcuate ridges (QT12-1c). In HyTES images, lava flow QTf4-3 appears to flow around two 10-20 m relief domes (QTfz) that were not mapped separately from QTf4-3. The color change in the image (Fig. 1) along lava flow QTf12-1c appears to correlate with greater amounts of lighter purple color in the east, so there is a quantifiable change. The change in the rock might be remnants of sedimentary deposits on top of the flow, or possibly the flow was altered slightly under sediment cover that has been mostly eroded away, so only field data can resolve the reasons for the color variations. Once the processes are identified, they might be applicable to other deposits. HyTES images depict significant variations in the Quaternary sedimentary deposits (for example, Q4 and Q2b), although most are not separated in this cursory summary.

Conclusions

Comparison of the geologic map, aerial photographs, and HyTES L1 BT RGB and L2 LST images, indicate that many map units and depositional features are identifiable in all three data sets. The geologic map provides an important generalized spatial correlation of related deposits that is based on field studies. Aerial photographs and HyTES images provide higher resolution of spatial, shape, and tonal information for visual (qualitative) interpretation of features. This collocation on aerial photographs and HyTES images builds confidence in the visible identification and confirms the potential use of spectral characteristics of specific rocks or features. Digital National Agriculture Imagery Program (NAIP) images have three visible light bands (Red, Green, and Blue), and recent images have an additional near infrared band (NIR). Using NIR-G-B bands (NAIP bands 4, 2, 3, respectively) as RGB image channels results in false color images that typically accentuate differences in rocks (Buesch and Harvey, 2022). These four bands of NAIP data can provide a simple four-fold spectral characterization of rocks and sedimentary deposits. The advantage of the hyperspectral HyTES data for rock characterization and mapping, as with other data such as that supplied by Mako (Adams and others, 2022; Buesch and Harvey, 2022), is that there are many diagnostic spectral characteristics in the thermal infrared part of the electromagnetic spectrum that can be quantitatively used to identify rocks, alteration, and sediments and variations within these units.

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References


Something old, something new: what microfossils in archived cores and outcrop reveal about the paleohydrology of the Cronise lakes, Soda Lake, and “Lake Dumont,” lower Mojave River corridor, southeastern California, USA

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The evolution of the Mojave River and various paleolakes (e.g., Harper, Manix, Mojave; Fig. 1) along its course have been studied for over a century, yielding important insights into the integration and paleohydrology of the Mojave River watershed (e.g., Enzel et al., 2003; Wells et al., 2003; Reheis et al., 2012; Garcia et al., 2014). Here we present new preliminary microfossil and ostracode-based oxygen isotope results from archived cores from the East and West Cronise basins and from Soda Lake (Lake Mojave). We update previous Salt Springs basin (“Lake Dumont”) interpretations with a new age constraint and new ostracode-based oxygen isotope results. Our results provide additional insights into the characteristics of three late Pleistocene and Holocene desert lake and wetland paleoenvironment archives along the lower reaches of the Mojave River.

East and West Cronise basins

Cores were taken from East and West Cronise basins (EC-1 [19-m-long] and WC-1 [37-m-long], respectively) in 1995 and sampled for microfossil analysis in approximately 2005. Those results have never been published. To our knowledge, EC-1 and WC-1 remain archived at the Desert Research Institute, Reno, NV.

The upper 13.5 m of core EC-1 contains ephemeral, intermittent, and perennial lake sediments spanning the Holocene, based on three previously published AMS age constraints and sedimentological interpretations (Wells and Anderson, 1998). The upper 2.2 m of core WC-1 contain intermittent and perennial lake sediments that were deposited approximately 11.3 14C ka BP, based on a single previously published AMS age estimate (Wells and Anderson, 1998). New results show that the ostracode fauna from the lacustrine intervals in EC-1 and WC-1 is dominated by Limnocythere ceriotuberosa, similar to other Mojave River-fed paleolakes. Rare valves of Limnocythere bradburyi.
occur in late Holocene sediments near the top of core EC-1. Although *L. bradburyi* occurs in Pleistocene lake sediments at several locations in the southwestern U.S., the current range of *L. bradburyi* does not extend north out of Mexico into the continental U.S., and Holocene-aged valves are only known from one other location at Summer Lake, Oregon. Newly obtained oxygen isotope values (δ¹⁸O) from *L. ceriotuberosa* valves in EC-1 and WC-1 range from -3.6‰ to 15.9‰ VPD. Highest δ¹⁸O values are associated with the ephemeral mid- and late-Holocene lake cycles in core EC-1. The unexpectedly large range and abnormally high values suggest a combination of extreme evaporation and perhaps contributions from monsoon (or tropical) precipitation. Currently, the North American Monsoon does not extend into the Mojave Desert and Cronise basins watershed except only during the strongest monsoon years (Barron et al., 2012). However, regional lake, speleothem, and midden records suggest that the North American Monsoon was more active in the Mojave Desert at various times during the late Holocene (Metcalf et al., 2015).

**Soda Lake**

Soda Lake cores 1 and 4 were collected by the U.S. Geological Survey in the early 1950s (Muesig et al., 1957). Rick Forester (USGS-deceased) analyzed numerous microfossil samples from core 1 in the early 2000s, primarily from samples located 10–35 meters below the surface in a sequence of clay-rich sediments deposited by late Pleistocene Lake Mojave. The results of his study have never been published. Boxes containing numerous fragments of Soda Lake core 4 were located by Marith Reheis (USGS-retired) in 2008. Several samples from core 4, from seven to 20 meters below the surface, were taken for microfossil analysis and featured in an unpublished senior thesis at Northern Arizona University (NAU; Flagstaff, AZ) later that year.

In Soda Lake core 1, ostracode valves from approximately 25 m and 26 m below the surface were deposited 18.0 ¹⁴C ka BP and 18.8 ¹³C ka BP, respectively, based on previously published AMS age estimates (Reheis et al., 2015). Most of the core 1 sediments studied by R. Forester contain both *L. ceriotuberosa* and *L. bradburyi*, suggesting deposition during Lake Mojave I (≥17 to 13.5 ¹³C ka BP). In Soda Lake core 4, *Limnocythere ceriotuberosa* and *L. bradburyi* co-occur in only four of the stratigraphically lowest samples, suggesting deposition during Lake Mojave I. The remaining stratigraphically higher samples contain only *L. ceriotuberosa*, suggesting deposition during Lake Mojave II (13.4 to 10.9 ¹³C ka BP). These results are consistent with ostracode occurrences reported in more recent Soda Lake cores (Honke et al., 2019). Overall, δ¹⁸O values from all Soda Lake *L. ceriotuberosa* and *L. bradburyi* valves range from -3.0‰ to 4.7‰ VPD and from -2.4‰ to 5.3‰ VPD, respectively. Although there is considerable overlap, the difference in δ¹⁸O values between the two ostracode species is statistically significant (P value = 0.003). The slightly higher δ¹⁸O values in *L. bradburyi* valves could be consistent with a slight preference for calcification during summer months in more evaporatively evolved lake water. In Soda Lake core 4, the uppermost samples containing only *L. ceriotuberosa* (Lake Mojave II) consistently yield a portion of ostracode valves with lower δ¹⁸O values than found in the stratigraphically lower, mixed assemblage samples (Lake Mojave I). This might suggest that the inflow of the Mojave River during Lake Mojave II was more voluminous and perhaps had a lower δ¹⁸O value than during Mojave I. Continued incision of Afton Canyon (Reheis and Redwine, 2008) and perhaps more efficient draining of upstream lakes (e.g., Lake Coyote; Miller et al., 2018) would have reduced the amount of time Mojave River water might have been impounded and evaporated upstream before reaching Lake Mojave II. Or, perhaps decreasing accommodation space caused Lake Mojave II to overflow more frequently than Lake Mojave I, increasing its potential for generating *L. ceriotuberosa* valves with lower δ¹⁸O values than was typically possible during Lake Mojave I.

**Salt Springs basin and “Lake Dumont”**

The Salt Springs basin (hereafter "Dumont basin") was cored in 1995 (DU-1 and DU-2). To our knowledge, these cores remain archived at NAU. Core DU-2 (16-m-long) and several Dumont basin outcrop locations were sampled for their microfossil content in 2005. The results of that endeavor were featured in a Friends of the Pleistocene field trip guide (Bright and Anderson, 2007), but are updated within this short paper.

Previously published AMS estimates from 10.8 m and 0.76 m depth in core DU-2 suggest deposition prior to 27.5 ¹⁴C ka BP up to approximately 18.2 ¹⁴C ka BP (Anderson and Wells, 2003). Surficial “Lake Dumont” sediments abutting bedrock at approximately 177 m above sea level along the eastern flank of the northern Salt Spring Hills (i.e., located north of Hwy. 127) contain snail shell fragments that yielded a new direct carbonate AMS age estimate of 12.0 ¹³C ka BP. An aliquot of *Candona patzucaro* valves from an outcrop of olive-green sediments exposed in an arroyo wall several hundred meters basinward from the northern Salt Spring Hills yielded a direct carbonate Fraction Modern Carbon value that was lower than the procedural blank. The age of these ostracode valves is considered indeterminate. Dumont basin outcrops and the entirety of core DU-2 contain indistinguishable ostracode faunas that are incompatible with deposition in a Mojave River-fed lake (Bright and Anderson, 2007). The key ostracodes, *Cyprideis beaconensis* (a marine ostracode that inhabits coastal areas of western North America) and *Limnocythere staplini*, both inhabit low carbonate alkalinity-to-calcium (ALK/Ca) waters. Both species would find the elevated salinity and high ALK/Ca chemistry of a terminal Mojave River-fed lake to be intolerable. New δ¹⁸O values from
co-occurring *C. beaconensis* and *C. patzucaro* valves in two outcrop samples of the aforementioned olive-green sediments range from -8.6‰ to -4.8‰ VPDB and from -5.4‰ to 3.8‰ VPDB, respectively. *Cyprideis beaconensis* valves associated with the newly dated snail fragments yielded δ¹⁸O values of -13.4‰ and -12.1‰ VPDB. Most of the ostracode valves analyzed from "Lake Dumont" yield δ¹⁸O values that are several per mil lower than ostracode valves from late Pleistocene Lake Mojave sediments and Holocene Cronise basin lake sediments. The low ostracode δ¹⁸O values suggest calcification in spring discharge (e.g., wetland) rather than in a lake sustained by overflow from Lake Mojave. The combined ostracode fauna and ostracode δ¹⁸O values from "Lake Dumont" are incompatible with a Mojave River-fed lake interpretation. Some of the Dumont basin sediments predate Lake Mojave and thus, could not have been deposited during periods of overflow from that lake. Dumont basin sediments that are coeval with upstream Lake Mojave still lack the appropriate Mojave River-derived ostracode fauna and lack elevated lacustrine δ¹⁸O values. Our interpretation generates an intriguing question over whether or not the Mojave River reached Dumont basin during the late Pleistocene and contributed water to Lake Manley in Death Valley (e.g., Anderson and Wells, 2003).

**Summary**

This study highlights how archived cores can provide a wealth of paleoenvironmental information at a small fraction of the cost and effort of recovering new core material.

The combination of ostracode faunal interpretations and ostracode δ¹⁸O values clearly delineate perennial lakes (Lake Mojave), ephemeral lakes (Cronise lakes), and wetland/spring deposits ("Lake Dumont") along the lower Mojave River corridor. Ostracode δ¹⁸O values reveal pronounced differences in hydrologic characteristics among the three study locations. At Soda Lake basin, perennial Lakes Mojave I and II were both supported by high ALK/Ca Mojave River water and were isotopically evolved, but with a higher degree of isotopic variability during Lake Mojave II. Ephemeral mid- and late-Holocene lake cycles in the Cronise basins were supported by high ALK/Ca Mojave River water but experienced pronounced evaporation. The abnormally high ostracode δ¹⁸O values (> 10‰ VPDB) in the upper part of core EC-1 could suggest that monsoon (or tropical) moisture delivery to the eastern Mojave Desert was higher during the late Holocene than it is today. And finally, the ostracode fauna and ostracode δ¹⁸O values from Late Pleistocene "lake" sediments in Dumont basin are strikingly different from all other upstream Mojave River-fed lake deposits. Between roughly 27 ¹⁴C ka BP and 12 ¹⁴C ka BP, Dumont basin appears to have consistently held wetlands supported by low ALK/Ca groundwater with low δ¹⁸O values. The leading explanations are that Dumont basin never held a Mojave River-fed lake, or evidence for a Mojave River-fed lake still awaits discovery, or perhaps the necessary evidence has been eroded way over the past several millennia.

**References**


Post-Early Miocene silicic volcanism in the northern Mojave Desert, California

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ABSTRACT—Silicic volcanism that postdates widespread early Miocene volcanism in the Mojave Desert has not been extensively studied. We compiled age, petrographic, and geochemical data for volcanic rocks in a wide swath of the desert south of the Garlock fault using an age threshold of post-18.8 Ma, approximately the limit of the earlier Miocene volcanism as marked by the eruption of the widespread Peach Spring Tuff. In addition to the well-known young basaltic volcanic centers not considered in this paper, several dozen silicic volcanic edifices are known or likely to be younger than 18.8 Ma. Several examples of rhyolite tuffs and basalt lava in the middle Miocene basin occur in sequences of the Barstow Formation and its correlatives. Dacite domes are common in the Calico Mountains, dated at ~17 Ma, and similar, but mostly undated, domes are scattered nearby in the Barstow area and east of the Calico Mountains. North of Barstow, chains of rhyolite domes and scattered dacite domes are known. A few of these domes and flows are dated in the range of 13-7 Ma. Farther north, the Lava Mountains have several volcanic sequences from 12 to 7 Ma that range in composition from basalt to rhyolite. Farther east and west are more rhyolite and dacite domes, in general undated, as well as the extensive ~17.8 Ma Woods Mountains rhyolite center. Sparse geochemical data for the silicic rocks indicate distinct rhyolite and dacite groups, and rare andesite. Understanding of these potentially young silicic volcanic rocks is hampered by poor age control and geochemical data, but more study could enhance the understanding of the origins of volcanism in the Mojave Desert.

Introduction
The north-central Mojave Desert (south of the Garlock fault and north of the Transverse Ranges, Fig. 1) and adjacent areas experienced volcanism and continental extension in scattered locations during the early Miocene. This event varied in space and time, but across much of the Mojave Desert, from Barstow east to the lower Colorado River corridor, it appears to have been mainly 21 to 19 Ma, with the Peach Spring Tuff (18.8 Ma) serving as a capping flow that post-dated extension in many locations. Examples include tilted volcanic sections in the Bristol Mountains and Lava Hills in the east (Miller, 1994), the Cady Mountains and Alvord Mountain (Byers, 1960; Woodburne et al., 1982), and the Barstow area (e.g., Fillmore and Walker, 1996). Volcanism was younger in the northern Colorado River corridor (e.g., Faulds et al., 2001) and in the Eagle Crags area of the north-central Mojave Desert. Volcanic sequences are also known in parts of the western Mojave Desert. However, many volcanic rocks are only poorly dated as early Miocene in age as part of the Tropico Group. It is widely accepted that silicic volcanism ceased or diminished significantly after about 19 Ma.

This paper addresses volcanism after the Peach Spring Tuff eruption by cataloging isolated lava flows and domes in the Mojave Desert as well as tuffs and lavas of the Barstow Formation and its correlatives. The rocks are not dated well, ranging through the middle and late Miocene and possibly into the Pliocene, and volcanism was less voluminous than pre-Peach Spring Tuff sequences. The silicic volcanism is relevant to several questions, and with more study these can be addressed: How do spatial and temporal patterns relate to areas of earlier volcanism? What is the source of the silicic magmas and how do compositions compare to earlier rocks? How does the silicic volcanism relate to crustal composition and tectonic setting?

Setting
This paper treats a broad area of the northern Mojave Desert, framed by the San Andreas and Garlock faults and extending east nearly to the Colorado River (Fig. 1). Extensive work in the Colorado River corridor has established that extension and volcanism progressed northward with time, from late Oligocene near Mexico to late Miocene near Las Vegas. West of the Mojave Desert, the plate-boundary San Andreas fault formed and lengthened with time, shutting down subduction and dramatically changing the crustal stress (Atwater, 1970).

Although a few tectonic syntheses of the northern Mojave area have been written (e.g., Glazner et al., 2002), none specifically focused on middle Miocene and younger volcanic rocks. These syntheses have argued that
much of the area underwent early Miocene extension, forming extensional basins dominated by volcanic and volcaniclastic rocks. Examples of the 21-19 Ma volcanism and normal fault-bounded basins range from the Bristol Mountains and Cady Mountains in the east (Miller, 1994) to the Newberry Mountains south of Barstow (Dokka et al., 1991) and several mountains north of Barstow (Fillmore and Walker, 1996). A younger volcanic system, the Eagle Crags field, lies in China Lake Naval Weapons Station and adjacent Fort Irwin; volcanic rocks range from 21 to 16 Ma, and extension is minor or absent (Schermer et al., 1996; Sabin et al., 1994; Buesch et al., 2018). Many authors argue that after early Miocene volcanism and extensional tectonics waned, a period of subdued tectonics led to the development of broad shallow basins along an east-trending belt from the Boron area to the Cady Mountains (e.g., Woodburne, 1991; Dokka, et al., 1991; Fillmore and Walker, 1996). The basins accumulated lacustrine and fluvial sediment, coarse-grained in a few areas, and have few volcanic rocks; the sedimentary sequences are known as the Tropico Group, Barstow Formation, and upper Hector Formation (Woodburne et al., 1982; Woodburne, 1991). This simple picture of post-tectonic basins has proven awkward with the clarification of times of initiation within several basins containing the Barstow Formation. Initiation of lacustrine sedimentation predates the Peach Spring Tuff in some basins and is 2 to 3 million years younger in others (Miller et al., 2010). Basin sedimentation was disrupted by initiating strike-slip tectonics associated with the eastern California shear zone ~10-11 Ma (Nuriel et al., 2019). This strike-slip regime has continued to present.

The existing paradigm for Miocene stratigraphy is that coarse-grained volcanic and volcaniclastic rocks belonging to the Pickhandle Formation and its correlates are largely pre-19 Ma and are syntectonic with extensional basin development and that younger, predominantly fine-grained lacustrine facies rocks of the Barstow and upper Hector formations are post-extensional and represent the gradual infilling of the previously formed extensional basins (e.g., Fillmore and Walker, 1996). Our view is that this distinction between pre-19 Ma coarse-grained/syntectonic and post-19 Ma fine-grained post-tectonic breaks down in detail. Both fine-grained and coarse-grained facies were present throughout the Miocene sedimentary history. Rather, the distinction is between volcanic-dominated basins before ~19 Ma and volcanic-poor basins afterward. With those caveats in mind, this paper focuses on the scattered silicic volcanism that is probably younger than the Peach Spring Tuff (18.8 Ma) and appears to post-date most extensional...
tectonism in the region, and is distinctly younger than the volcanic-dominated sequences broadly correlative with the Pickhandle Formation (Woodburne, 1991).

Assigning equivalent syn- and post-tectonic times to strata of the Tropico Group (Dibblee, 1960b; 1963), taken as broadly correlative to the Pickhandle Formation, in the western Mojave Desert is similarly problematic, especially given the paucity of reliable dates. The timing of extension and volcanism is diachronous in the Colorado River corridor, bearing some similarities to the area farther west that we discuss. Widespread thick sequences of volcanic rocks of the Eagle Crags–Fort Irwin area extend to a younger timeframe than closer to Barstow. In the Fort Irwin area, a thick Miocene volcanic-dominated section ranges from about 21 to 16 Ma (Scherrner et al., 1996; Buesch et al., 2018). The younger, largely sedimentary, section is about 16 to 10 Ma at Fort Irwin. Farther west, the Eagle Crags volcanic field appears to have a similar range of ages for volcanism, as well as a few volcanic centers that postdate 15.5 Ma (Sabin, 1994; Buesch et al., 2018), as do the Lava Mountains adjacent to the Garlock fault (Smith et al., 2002; Andrew et al., 2014, Andrew and Walker, 2017).

**Methods**

We compiled geochemical data from published sources and new data from U.S. Geological Survey contract labs in order to create summary plots. As a result, there is variation among methods. In general, major elements, reported herein as oxides in weight percent, with volatile-free normalized (summed to 100%) values, were determined by wavelength dispersed X-ray diffraction (XRF). Trace elements were determined by inductively coupled plasma atomic emission spectroscopy and are reported in parts per million (ppm).

Chronologic control is critical for assessing young silicic volcanism, and we report ages derived from published and new work that uses several methods, each with their inherent uncertainties and calibration issues. We have not attempted to normalize all ages to a single age standard and note that it may not be possible to directly compare the ages obtained in different labs or by different methods. We report all age uncertainties at 2σ, but emphasize that in many instances, these are simply an approximation of the analytical uncertainties but that many of the samples yielded complicated results (e.g., disturbed 40Ar/39Ar age spectra, multiple generations of zircons). These assigned uncertainties may severely underestimate the true “geological” uncertainty in the ages.

**40Ar/39Ar analytical Methods.** Standard density and magnetic separation techniques and handpicking were used to generate purified separates of plagioclase, biotite, K-feldspar, hornblende, and holocrystalline groundmass (for basalt) from the various samples. All samples were analyzed in the 40Ar/39Ar geochronology laboratory at UC Santa Barbara by incremental heating in a Staudacher-type resistance furnace. Isotopic analyses were obtained on a Mass Analyzer Products 216 mass spectrometer, using the general procedures and system described by Gans (1997). The flux monitor used for all irradiations was Taylor Creek Rhyolite with an assigned age of 27.92 Ma (Dalrymple and Duffield, 1988). For comparison, we obtained an age of ~27.75 Ma on Fish Canyon Tuff sanidine (another widely used standard). It is important to note that direct comparison of the U-Pb ages to the 40Ar/39Ar ages obtained in this study depends critically on the assumed ages of standards used in the analyses. Our assumed age of 27.92 Ma for the Taylor Creek Rhyolite sanidine 40Ar/39Ar flux monitor yields ages that are approximately 1.5% younger than ages obtained from U-Pb dating on equivalent units (e.g., 27.75 Ma versus the 28.20 Ma for Fish Canyon Tuff) (Kuiper et al., 2008). Thus, the 40Ar/39Ar ages reported here from the 15 to 19 Ma volcanic rocks should be increased by ~0.24 to 0.30 Ma to make them directly comparable to the U-Pb ages. In many cases, this difference lies within the range of analytical uncertainty. All analytical errors given for our estimated (preferred) ages as reported throughout the text and in Table 1 are ± 2σ (95% confidence). Analyses ranged from total fusion experiments on single grains to multiple-step incremental heating experiments. Data reported in Table 1 distill those samples for which replicate splits and/or multiple minerals were analyzed to one value.

**U-Pb analytical methods.** Zircon grains were separated using standard crushing and heavy liquid techniques, mounted in epoxy, polished, and imaged by cathodoluminescence prior to analytical sessions. Isotopic analyses for U-Pb geochronology and trace element geochemistry were performed simultaneously by secondary ion mass spectrometry using the Stanford-USGS SHRIMP-RG ion microprobe housed at Stanford University, USA. Analyses followed the analytical protocol and data acquisition conditions described by Watts et al. (2016). Raw data were reduced using the SQUID2 software version 2.51 (Ludwig, 2009) and employed a Pb/U-UO/U calibration (Williams, 1998) keyed to the R33 zircon standard with an age of 419 Ma (Black et al., 2008). Reported dates and weighted means were calculated using Isoplot version 3.76 (Vermeesch, 2018). Trace element concentrations were calculated from secondary ion yields normalized to 90Zr. 16O-corrected to the yield from concurrent analyses of MAD-1 (and MAD-559) zircon standards (Coble et al., 2018).

**Volcanic rocks similar in age to the Peach Spring Tuff**

The Peach Spring Tuff is a regional outflow sheet that erupted from a center within the Colorado River corridor (Ferguson et al., 2013) at 18.8 Ma based on 40Ar/39Ar dating of sanidine. It provides a regional marker associated with a significant break in stratigraphy in many places. For instance, it angularly overlies tilted rocks of
small extensional basins in the Bristol Mountains and Lava Hills (Miller, 1994), tilted volcanic rocks in the Newberry Mountains (Dokka et al., 1991), tilted rocks in Alvord Mountain (Byers, 1960), and complexly faulted volcanic rocks of Daggett Ridge (Dokka et al., 1991; Miller et al., 2010). In the Daggett Ridge area (Fig. 2), it is found within the lower part of the Barstow Formation.

In some places, volcanic rocks occur at a similar stratigraphic position to the Peach Spring Tuff, but they have not been dated in detail. In the Hector Formation of the Cady Mountains south of Afton Canyon, the Peach Spring Tuff overlies a section that includes thick white tuffs (Moseley, 1978). At Alvord Mountain, Byers (1960) designated the thin Spanish Canyon Formation as a unit that overlies tilted conglomerate of the Clews Formation but underlies the Barstow Formation. The Spanish Canyon Formation includes several rhyolite tuffs and basalts and sandstone (Buesch et al., 2013) including the Peach Spring Tuff in the upper part of this section. Approximately nine km northeast of Alvord Mountain, a similar section of interbedded tuffs and basalt/andesite may correlate with the Spanish Canyon Formation.

Volcanic rocks within or correlative to the Barstow Formation (19 to 13 Ma)

Several ash beds and dacitic domes are mapped within the Barstow Formation and correlative units in the central Mojave Desert, although volcanic rocks are sparse compared to the underlying Pickhandle Formation. The most extensive outcrops are in a complex of ~15 domes, the Yermo volcanic field in the southern Calico Mountains (Fig. 2), studied by Singleton and Gans (2008). The domes intrude and lie on Barstow Formation beds, and some Barstow beds lap onto dacite locally. The dacites are homogeneous and contain plagioclase, hornblende, and sparse biotite phenocrysts. Although some are potassically altered, less altered rocks plot within the dacite field on a LesBas (1986) total alkali silica diagram (Fig. 3A) with SiO$_2$ ranging from 65.2 to 67.7 wt. % and total alkalis about 6 to 7 wt. %. Potassically altered rocks plot as trachydacite, with total alkalis as great as 9 wt. %.

For these rocks, elevated K$_2$O is accompanied by decreased Na$_2$O and CaO. Dating by $^{40}$Ar/$^{39}$Ar methods by Singleton and Gans (2008) yielded ages for the domes (whole rock and plagioclase) from 16.8 ± 0.2 to 17.1 ± 0.1 Ma. The volcanic field is 25 x 10 km, although additional domes may lie south of the Calico Mountains beneath younger gravels. No domes of this field are known within Barstow Formation exposures east of the mountains.

Southwest of the Calico Mountains in the southern Mitchel Range (Fig. 2), Van Pelt (2011) identified and dated several domes that intruded the Pickhandle or Barstow rocks and deformed and baked their wallrocks. A biotite dacite dome (0.25 km$^2$) that intruded the Barstow Formation consists of reddish flow banded, phenocryst-rich (25%) rock. It was dated ($^{40}$Ar/$^{39}$Ar; plagioclase) at 16.98 ± 0.10 Ma. The dome is similar in age to the domes of the Yermo volcanic field but differs mineralogically in that biotite is much more common than hornblende. The largest dome, Elephant Mountain (0.55 km$^2$), is a composite of several hornblende dacite domes intruding the Pickhandle Formation and was dated ($^{40}$Ar/$^{39}$Ar; plagioclase) as 16.98 ± 0.10 Ma on plagioclase and 17.21 ± 0.10 Ma on biotite. The dome is similar in age to the domes of the Yermo volcanic field but differs mineralogically in that biotite is much more common than hornblende.
similar in composition to, Elephant Mountain. It yielded a plagioclase date (Ar/Ar) of 18.49 ± 0.32 Ma (Van Pelt, 2011).

In the Barstow area, we mapped several domes. All except one small northern dome are exposed in hills surrounded and partially buried by Pliocene or early Pleistocene alluvial gravels (Dibblee, 1960a). Some of these domes are similar mineralogically to those of the Yermo volcanic field but are small (0.1-0.2 km²) and generally widely spaced. If they constitute a fringe of the Yermo field, that field may encompass several dozen domes. The northern dacite dome is dated at 19.60 ± 0.10 Ma (02PGMJ-13; Table 1) and is therefore older than the Yermo volcanic field. The western dome is compound and much larger than others at 0.7 km². The eastern two domes constitute a silicified and argillically altered complex of small domes, flows, and tuff; the lavas are reddish-brown plagioclase-sanidine-quartz-biotite rhyolite with platy structure.

**Ash beds within the Barstow Formation.** Ash beds are present in outcrops of the Barstow Formation and correlative units at several locations between the Gravel Hills and Alvord Mountain (Fig. 1). In the general area of the Yermo volcanic field (Fig. 2), the Barstow Formation includes several tuffaceous beds. In the Mud Hills several ash beds have been dated, providing a relatively complete chronology of the Barstow (MacFadden et al., 1990; Woodburne et al., 1990). Although most tuffs occur as ashy sediment or thin beds and may represent reworked fallout ash from distant eruptive sources, the Skyline Tuff ranges from 2 to 5 m thick (Dibblee, 1960a) which is 5-10 times thicker than others we have studied in this formation. We dated this zeolitized tuff at the Mud Hills (sample M12Cl-905, Miller and Vazquez, 2022 and Table 2), where it lies above the Valley View Tuff (15.3 ± 0.2 Ma) and below the “Dated tuff” (14.8 ± 0.1 Ma; both ages from MacFadden et al., 1990). We obtained a 15.2 ± 0.1 Ma age by U-Pb methods on zircon (Fig. 4; Miller and Vazquez, 2022 and Table 2, combined data from two samples). All grains had a similar appearance and yielded similar ages, and thus probably represent a primary fallout deposit. The tuff is composed of two layers at the sample site: a lower layered tuff with zeolitized glass plus minor biotite and feldspar, and a massive upper unit with similar composition and soft sedimentary structures indicating reworking. Lower in the Barstow section, the Rak Tuff, sampled from a tuffaceous interval in siltstone and sandstone, yielded equivocal Ar results. The Ar/Ar age assigned to this ashy interval is 16.3 ± 0.3 Ma based on three heating experiments on somewhat altered biotite and several single-grain fusions, but the data indicate ages ranging from 20.4 to 14.6 Ma (MacFadden et al., 1990). The results are not easily reconciled, and MacFadden et al. (1990) selected the mean of the three heating experiments as the best age. Our zircon work yielded a U-Pb age of
20.4 ± 0.2 Ma (sample M12CL-902, Miller and Vazquez, 2022 and Table 2) on a coherent group of 10 grains. A similar result is obtained if all 15 Cenozoic grains are grouped. The zircon grains behave as a simple coherent population, and we interpret them as reworked although from a single eruption. The zircon date fails to clarify the uncertain assignment of an age to this interval in the Mud Hills, although it is similar to the oldest fusion ages for associated biotite grains. Alternatively, the youngest biotite grains may best approximate the eruption and depositional age of the ash bed. We also dated by 40 Ar/39 Ar a tuff bed high in the section within the syncline at Rainbow Basin, where it overlies the Skyline and "Dated" tuffs. The results (RB-5, Miller and Vazquez, 2022 and Table 1) yield an age of 13.44 ± 0.05 Ma, similar to the age given by MacFadden et al. (1990) for the Lapilli Tuff.

We dated a yellowish, thick aphyric altered ash bed east of the Mud Hills, which lies in a gypsiferous section of Barstow mudstones. A similar ash bed lies 20 m higher in the section. The beds can be traced west to a position that indicates possible continuity with the Skyline Tuff. The yellow tuff (sample M11NS-2409, Table 2) had sparse zircon, of which fewer than half were younger than Mesozoic. The four young zircons yielded an age of 16.5 ± 0.2 Ma. This tuff apparently lies lower in the section than the Oreodont Tuff (15.8 Ma) in the Mud Hills (MacFadden et al., 1990) and illustrates the rapid facies changes in the Barstow basin in addition to the locally thick tuffaceous deposits.

A U-Pb age slightly older than the Skyline Tuff was obtained for an altered 50-80-cm thick white ash bed in the Yermo Hills (sample M9NS-3193, Miller and Vazquez, 2022 and Table 2). The Upper Emerald ash bed has sanidine phenocrysts as well as incipient opalization. The opal displays a breccia fracture pattern. This sample yielded a U-Pb age of 15.2 ± 0.2 Ma, which revises a previously reported age of 15.4 ± 0.2 Ma (Miller et al., 2010; Miller and Vazquez, 2022). Zircon grains for this sample are distinctive: they are broken, long and thin in shape with aspect ratios of 4:1 to 7:1, and they have distinct oscillatory zoning and numerous melt inclusions. In contrast, zircon of the similarly aged Skyline Tuff is stubby (1:1 to 2:1) and less distinctly zoned and has few melt inclusions. These observations support the interpretation that the Upper Emerald ash is distinct from the Skyline Tuff despite similar U-Pb ages. Below sample M9NS-3193 in the Yermo Hills are several white to green ash beds, mostly aphyric, that can be traced laterally for over 100 m in the mudstones that enclose them. The lowest phenocryst-bearing ash in the section is ~55 m below the 15.2 Ma Upper Emerald ash; it is a green Emerald ash dated at 16.8 ± 0.2 Ma (M10NS-17, Miller and Vazquez, 2022 and Table 2). This ash yielded more zircons forming a coherent population than most ash beds we have dated. A few Mesozoic grains were among the Miocene grains.

Several other ash beds occur in the Barstow Formation, and we attempted to date some (Miller et al., 2010; 2013a), but few had sufficient primary zircon to allow a confident age assignment (Miller and Vazquez, 2022). We consider most of their ages to be poorly defined. In the Calico Mountains (Fig. 2), the Barstow Formation is constrained by dates on underlying dacite lava of 19.1 Ma, and much of the section underlies the ~17.0 Ma dacite of the Yermo volcanic field (Singleton and Gans, 2008). These workers noted that Barstow deposition elsewhere continued well after the dacite eruptions, and Miller et al. (2013b) demonstrated that near the dacite domes in the Calico Mountains, Barstow deposition continued after dacite eruptions. We attempted to add chronologic control in three places in the Calico Mountains: low in the section, where an altered ash bed occurs, higher in the section and along strike with dacite breccia that lies on lacustrine beds, and in a fault-bounded block at Cemetery Ridge, west of the Calico Ghost town and south of a major strand of the Calico fault. The low ash bed (Lower Calico ash M10NS-152, Miller and Vazquez, 2022 and Table 2) yielded just eight zircons that might
Figure 3. Geochemistry data for volcanic rocks younger than ~18.8 Ma at the latitude of Barstow and younger than ~16 Ma farther north in the Fort Irwin-Eagle Crags area. A. Total alkali-silica plot of volcanic rocks (Le Bas et al., 1986) discussed in this report along with fields for Pliocene basaltic centers and Early Miocene dacite of Eagles Crag field at Fort Irwin. B. Modified alkali lime index plot (Frost et al., 2001). C. Plots of trace elements used to discriminate tectonic environments (Pearce et al., 1984).

be primary, with a date of 17.6 ± 0.2 Ma. The sample higher in the section (Red-Green ash, RE-GR, Miller and Vazquez, 2022 and Table 2) yielded a large coherent set of zircon grains of 18.3 ± 0.1 Ma, an unexpectedly old result. Evidently, unidentified structure in this area of the Calico Mountains brings older parts of the section to a higher position. The fault-bounded block at Cemetery Ridge (Fig. 2) has two prominent white ash beds within limestone, sandstone, and mudstone, one of which yielded sparse young zircons (Cemetery Ridge ash, M10NS-013, Miller and Vazquez, 2022 and Table 2) among a large population of Mesozoic and older zircon. Of the young zircons, only three are possibly primary. These yielded a combined age of 18.2 ± 0.3 Ma. Interestingly, the three zircons have trace element characteristics like those in the Peach Spring Tuff (Miller et al., 2010) including high Th/U ratios and high values of Ce. Furthermore, zircons have bright, low-U rims in cathodoluminescence similar to those for the Peach Spring. The ash bed contains sparse biotite, feldspar, and quartz, and appears to be reworked.

The Lime ash, sampled at the top of the Barstow Formation at Lime Hill (Fig. 2), yielded an age of 15.3 ± 0.3 Ma on seven grains forming a coherent population. Overlying gravels mark a gentle angular unconformity; evidently, part of the upper Barstow was eroded.

**Correlation and origin of ash beds.** The zircon dating of ash beds in the Barstow yielded several similar ages (Fig. 4), raising the possibility that ash beds are correlative in some cases. We described above the case against the correlation of the 15.1 Ma Skyline Tuff with the Upper Emerald ash (sample M09NS-3193 in the Yermo Hills on the basis of zircon appearance. The Yellow tuff east of the Mud Hills (sample M11NS-2409), the Emerald ash in the Yermo Hills (sample M10NS-017), and the lime ash at Lime Hill (sample M10NS-452) also are broadly similar in age at ~15.7-16.5 Ma. Zircon grains in these samples were analyzed for trace elements during the dating experiments (Miller and Vazquez, 2022). All yielded similar results, which can be interpreted to indicate eruptions in areas with similar crustal columns with which the values are correlated. The data do not preclude but do not require a correlation of the ash beds.
Zircon trace elements can be used to evaluate crustal sources because the erupting magma interacted with crust, which is heterogeneous (e.g., Riggs et al., 2012) and offer a method to discriminate among the ash beds described above. We compiled zircon trace element data for several possible volcanic sources: the local Barstow area (Stone et al., 2019), the Eagle Crags volcanic field west of Fort Irwin (Sabin et al., 1994; Buesch et al., 2018; data for intrusive rocks from Barth et al., 2016), the Baker area (Barth et al., 2016), the northern Colorado River corridor, which has abundant middle Miocene plutons and volcanic rocks (Colombini et al., 2011; Clairborne et al., 2010), and the Yellowstone hotspot eruptions (Matthews et al., 2015). We compared trace element compositions of these sources with zircons from ash beds that yielded more than 5 grains. The data in some cases are strongly suggestive of eruptive sources. For example, in the plot of Th/U vs. Ce (Fig. 5A), many values for ash beds plot in an area in the lower-left where several potential source areas overlap, but the Upper Emerald ash and Emerald ash are strongly similar to the Colorado River corridor magmas with their high Th/U and Ce values. This correlation is supported by Y and La/Yb plots (Fig. 5D, C). The Lower Calico ash (“Little Borate tuff” of Miller et al., 2013a) and Lime ash are strongly similar to the Yellowstone magmas in having low Th/U values over a narrow range and variably high values of Y and Hf. This correlation is further supported by observations that unaltered glass shards in the Lime ash bed are chemically correlated with Yellowstone hotspot magmas, although tephrochronology correlations are inconclusive (Miller et al., 2013a). M.E. Perkins (University of Utah, written communication, 2014) examined this tephra and correlated it with the Cougar Point III ash, dated at 12.83 ± 0.03 Ma. No strong correlations to Mojave Desert source areas west of the Colorado River were found, although more thorough characterization of these possible sources is needed. Even so, it follows that many of the ash sources were distant from the Barstow area.

Other dated rocks in the Barstow area. West of the Calico Mountains a thick sequence of gently west-dipping heterolithic gravels underlies low hills south of the Calico fault and south of Old Fort Irwin Road. A bed of reworked ash was dated at 15.5 ± 0.2 Ma (40Ar/39Ar methods-04PGMJ-59, Table 1), indicating that the deposits are of Barstow age but are not lacustrine or distal piedmont facies as seen on the north side of the Calico fault (Fig. 2). A similar prominent white tuff bed within steeply dipping lacustrine beds of the Barstow Formation north of Daggett Ridge (02PGMJ-01, Table 1) is also dated at 15.5 ± 0.1 Ma. The ash bed is one of two, each ~50 cm thick and bearing biotite, plagioclase, quartz, and sandine crystals. The ash beds are very fine-grained but the abundance of crystals indicates a proximal source.

A proximal ash-flow tuff, perhaps rhyolitic, occurs in the upper part of a sandstone-dominated section near the town of Lenwood (Loughney et al., 2020). The tuff is lenticular in outcrop and characterized by reddish-gray tuff breccia with phenocrysts of sandine, plagioclase, quartz, and biotite in a glassy matrix. It is undated but probably within the ~19-13 Ma bounds of the Barstow Formation as defined at the Mud Hills because marker beds of the Barstow Formation are present (Loughney et al., 2020).

Elsewhere in lacustrine sections correlated with the Barstow Formation are occasional intercalated basalt flows that provide age control for the tuffs. At the Gravel Hills and Black Mountain (Fig. 2), Dibblee (1968) mapped basalt in and at the base of lacustrine beds of the middle Barstow Formation. The basal basalt was dated in the Black Mountains by Burke et al. (1982) by K-Ar at 16.5 ± 0.8 Ma. Our 40Ar/39Ar age of 17.70 ± 0.20 Ma (OM30ded03, Table 1) for this olivine basalt, a pillow lava, is probably more reliable but displays evidence for both argon loss and reactor-induced recoil. A nearby olivine basalt is 18.70 ± 0.20 Ma (OM31ded03, Table 1); it approximately dates the base of the Barstow Formation in the Gravel Hills. Conglomerate characterizes much of the Barstow Formation in the Gravel Hills, but a middle section of shale, siltstone, and fine sandstone contains tuff beds and a basalt interval. Loughney et al. (2020) described a basalt at the top of the fine-grained section, but Dibblee (1968)
Figure 5. Plots of zircon trace element patterns for tuff beds in the Barstow Formation. Envelopes labeled with place names show range of data for zircons from potential source areas. Large symbols for selected tuffs in Table 2. Reference data for known regions are given in the text.

described it as within the fine-grained lacustrine deposits; we obtained an age of 15.77 ± 0.10 Ma (02PGMJ-27, Table 1) for it. Tuffs lie within and above the lacustrine section. A tuff at the top of the lacustrine siltstone yielded an age of 14.00 ± 0.30 Ma (27ddGH05, Table 1). Higher in the sequence a tuff within a band of volcaniclastic rocks in the upper gravel of the Barstow Formation yielded an age of 13.25 ± 0.30 Ma (09ddGH05, Table 1). Based on its age, this upper tuff may correlate with the Lapilli Tuff of the Barstow Formation in the Mud Hills.

At Alvord Mountain (Fig. 6), two (but locally as many as three) basalt units are separated by thin-bedded arkosic sandstone. They were mapped by Byers (1960) in the middle of a unit of thin-bedded fluvial and lacustrine deposits that he correlated with the Barstow Formation. The basalt contains fine to medium grains of olivine, plagioclase, and pyroxene. Basalt units are composed of several flows and lie on reddened sediment. Swisher (1992) dated (K-Ar whole rock) the upper and lower basalt at 16.6 ± 0.4 and 16.5 ± 0.1 Ma, respectively. Chemically (Fig. 3; SiO₂ 50.2 wt %), it is lower in alkalis than the group of younger (<7 Ma) alkali basalts that form prominent edifices across the Mojave Desert (Pliocene basalt, Fig. 3).

Much of the middle and upper Barstow Formation at Alvord Mountain is tuffaceous, with pumice scattered in sandstone and slightly ashy beds in a few places, but distinct tuff beds are uncommon. Woodburne et al. (1982) reported K-Ar dates of 13.8 and 14.0 Ma on biotite and plagioclase from a white tuff in the “middle part of the section” at Alvord Mountain. Swisher (1992) subsequently reported a K-Ar date of 14.38 Ma, presumably for the same unit. We collected a hornblende rhyolite tephra high in the Barstow Formation in the Mud Hills. At Alvord Mountain (Fig. 6), two (but locally as many as three) basalt units are separated by thin-bedded arkosic sandstone. They were mapped by Byers (1960) in the middle of a unit of thin-bedded fluvial and lacustrine deposits that he correlated with the Barstow Formation. The basalt contains fine to medium grains of olivine, plagioclase, and pyroxene. Basalt units are composed of several flows and lie on reddened sediment. Swisher (1992) dated (K-Ar whole rock) the upper and lower basalt at 16.6 ± 0.4 and 16.5 ± 0.1 Ma, respectively. Chemically (Fig. 3; SiO₂ 50.2 wt %), it is lower in alkalis than the group of younger (<7 Ma) alkali basalts that form prominent edifices across the Mojave Desert (Pliocene basalt, Fig. 3).

Two thick beds composed of size-sorted (2-4 cm) pink pumice lie in similar Barstow Formation deposits in Fort Irwin 9 km northeast of Alvord Mountain (Fig. 6). The tuff is a pink sanidine-biotite-quartz rhyolite. The thick beds lack sediment and are suggestive of a nearby source. Sediment above the pumice beds contains much pumice and locally contains reworked tuff fragments with similar mineralogy. A pumice clast, probably from this location, was dated (K-Ar on biotite) by Swisher (1992) as the
“Cronese Tuff” at 12.48 and 12.62 Ma, which he combined as 12.6 ± 0.1 Ma. Outcrops of plagioclase-olivine basalt/andesite and interlayered altered white silicified rhyolite tuff lie about 1 km to the east and are of uncertain relation to the Barstow Formation. They may represent the basalt-and tuff-bearing Spanish Canyon Formation, a unit underlying the Barstow Formation at Alvord Mountain (Byers, 1960).

Volcanic rocks in the eastern Mojave Desert. Several early Miocene basins with basalt to rhyolite flows are known in the eastern Mojave Desert (e.g., Sherrod and Nielson, 1994). Other than the Colorado River corridor (not considered in this paper), younger silicic rocks are known only in the Woods Mountains (Fig. 1), where the remnants of a large caldera complex that formed thick outflow sheets over 600 km² occur (McCurry 1988; McCurry et al., 1995). The caldera was about 10 km wide, and eruptions were sufficiently energetic to eject basement lithic clasts as large as 20 m wide near the source. The K-rich rhyolite of this system was dated by $^{40}$Ar/$^{39}$Ar at about 18.5 to 17.6 Ma by McCurry et al. (1995) (but analytical data are not reported). Early eruptions, 18.5-17.8 Ma, were small-volume and widely distributed. Three major eruptions of high-silica rhyolite, 17.8-17.7 Ma, created widespread ignimbrites with total volume exceeding 80 km³, and potentially could be present as ash fallout beds over a wide area. The Woods Mountains volcanic event ended with multiple post-caldera silicic lava domes and flows from 17.7 to 17.6 Ma.

Isolated dacite and rhyolite domes and flows (post Barstow time?)

Near Black Mountain and across the Coolgardie Camp plateau (Fig. 2), several domes and assemblages of volcanic flows stand above the pediment cut into Mesozoic granitoids. A stack of ‘blocky andesite flows’ was dated by K-Ar (Burke et al., 1982) and termed the andesite of Murphys Well. Biotite in the lowest flow was dated at 13.5 ± 0.4 Ma, an age similar to ages for tuffs in the upper Barstow Formation. Based on our reconnaissance, many of the other hills in the area are underlain by dacite domes. All were termed ‘andesite’ by Dibblee (1968), which suggests that the andesite dated by Burke et al. (1982) was also dacite. McCulloh (1960) distinguished domes by composition, which we show in Figure 2. We studied a few of the domes, all hornblende dacite, for which geochemical data are reported in. The dacite is similar chemically to other dacites (Fig. 3) that are described below. Also present are thick rhyolite lava flows at the north margin of the plateau and east of the Superior Lakes. The flows are deeply eroded, with their bases exposed as coarse tuffaceous deposits lying on Mesozoic plutonic rock. This field of domes and flows is little-studied but may represent volcanism as young as ~13 Ma as indicated by the K-Ar date on one dome.
Several dacite domes are exposed east of Alvord Mountain in the Cronese Hills (Fig. 6). Walker et al. (1990) mapped two dome complexes, one 0.5 km² and the other 3 km², that lie on metamorphic rocks. The larger one is ~170 m tall and is a composite of hornblende and hornblende-biotite dacite domes and flows, with carapace breccia and block and ash flow exposed at the margins. The smaller is an elongate dome or composite dome of hornblende dacite. We mapped six more domes northwest of those mapped by Walker and one isolated dome to the southeast near I-15 and Cave Mountain (Fig. 6). The domes range from 0.5 to 0.02 km² in width and 10-70 m in height above the surrounding gravel. Most domes are mineralogically homogeneous, all bearing acicular hornblende and fine- to medium-grained plagioclase with minor to common quartz in an aphanitic or glassy gray to reddish matrix. The dome near Cave Mountain contains biotite as well as hornblende. The domes are phenocryst-rich, with estimated 20-45% phenocrysts, of which plagioclase composes about 80-90%. Geochemically, they are similar as well, with SiO₂ values of 65-68 wt.%, K₂O ~2.0-3.0 wt.%, and high Ba and Sr values (Fig. 3). Most domes consist of foliated and folded flow-banded rocks, and margins have breccia and local block and ash deposits. The bases of the domes we studied are not exposed and may lie on bedrock as they do in the Cronese Hills. The northern domes are surrounded by fluvial gravels that in one place have interbedded ash that was correlated by tephrochronology to a Mesquite ash collected in Death Valley at ~3.5 Ma (E. Wan, U.S. Geological Survey, written communication, 2018), which places an upper age limit on it and nearby domes surrounded by the same gravels. The dome close to Cave Mountain is in an area with gravel deposits older than ~5 Ma, based on tephrochronology, and the dome is likely older than the gravels, but relations are uncertain. Several of these domes exhibit a distinctive halo of surrounding younger gravels, which are generally un cemented. Within 10-30 m of the domes, calcite cements the gravels, sometimes forming dense calcareous conglomerate. In addition, a few of the domes bear a silicified zone ~1 m wide that involves both dacite and sedimentary rock. The effect is clearly spatially related to the dacite domes. We interpret the cementation as caused by groundwater interacting with the volcanic glass, an interpretation similar to that of Andrew et al. (2014) for similar relations in the Lava Mountains (Fig. 7) in the northern Mojave Desert.

Farther east, volcanic rocks lie on Mesozoic substrate high in the western part of the Soda Mountains (Fig. 6) and at lower elevations along the west and southwest flanks. These little-studied volcanic rocks are of unknown age but the flat-lying nature and lack of evidence for extensional basin deposits are consistent with an age...
younger than about 19 Ma. We examined a dark brown dacite on the west flank that evidently is a faulted remnant of a dome. It is a plagioclase-hornblende phenocryst-rich rock with coarse grains. Farther south near Razor (Fig. 6), Glazner (1990) reported chemistry on a plagioclase-biotite-hornblende dacite with trace hypersthene that has somewhat elevated MgO, similar to two domes in the Cronese Hills area but unlike domes in the Yermo volcanic field. A rhyolite dome and flow complex lie south of the dacite dome. Although altered, these rhyolites contain phenocrysts of plagioclase, sanidine, quartz, biotite, and hornblende. Additional work is needed to better understand the west Soda Mountains volcanic field.

North of the Grass Hills in the Grass Valley Wilderness Area, rhyolite outcrops form long, imposing ridges, and steep-sided, isolated buttes and mesas that rise to 250 m above the surrounding alluvial surface (Fig. 7). Most of these outcrops were mapped previously as Tertiary hypabyssal intrusions (Jennings et al., 1962). The rhyolites are light gray, brown, and pale red, and generally have < 10% total phenocrysts of sanidine, quartz, and plagioclase, all 2 mm or less in size. Biotite and hornblende are present locally in sparse amounts (Felger, 2020). The rhyolite generally is well-foliated and in direct contact with Mesozoic plutonic rocks, which show little to no weathering or decomposition at the contact. Black vitrophyre is present locally at the contact and is generally less than 1 m thick. Columnar joints are also present locally at the contact and are typically ~10 cm in diameter, < 1 m in length, and oriented subhorizontally, like stacks of cordwood. This suggests a near-vertical contact, such as would be associated with an intrusive body. Cogenetic volcaniclastic units and sedimentary rocks are not common. Where present, they are mainly grayish tuff and tuffaceous sandstone with abundant pumice, and large clasts of plutonic rock, suggestive of locally derived deposits. The morphology and physical characteristics of the outcrops suggest the rhyolites represent a mix of subvolcanic intrusions, domes, and flows.

Most of the rhyolite outcrops are strongly aligned along a northwest trend, suggestive of structural control. A nearly continuous chain of outcrops about 15 km long and up to 1 km wide cuts across the wilderness area trending ~300° to 320°. North of this chain of outcrops, in and just outside of the northwest corner of the wilderness area, several isolated domes trend ~300°. The westernmost of those domes has more abundant phenocrysts than the other Grass Valley rhyolites, with about 30% total phenocrysts of quartz, feldspar, biotite, and possibly hornblende. In addition, the phenocrysts are larger, with some feldspars approaching 1 cm in size. This string of domes also aligns with the long axis of an oblong-shaped dome on the southeast flank of Almond Mountain (~ 5 km to the northwest).

Preliminary geochemistry results show the Grass Valley rhyolites have SiO₂ of 71-79% (Fig. 3A). Rhyolites in the chain of outcrops inferred to be subvolcanic intrusions and domes are high silica with SiO₂ of 77-79%. In comparison, the phenocryst-rich dome to the northwest is a low-silica rhyolite (SiO₂ 71%) with alkali-calcic tendencies (Fig. 3C). An isolated tabular flow that forms a prominent flat-topped mesa in the southwest part of the wilderness area looks similar in hand sample to the outcrops in the chain but has slightly lower SiO₂ of 74%.

None of the rhyolites in the Grass Valley area have radiometric ages, but they are inferred to be correlative to the Opal Mountain volcanics mapped by Dibblee (1968) in the Grass Hills and Black Mountain area about 10 km to the south. A sample from the Grass Hills analyzed by Burke et al. (1982) yielded a whole rock K-Ar age of 18.9 ± 1.3 Ma. Unpublished ages (P.B. Gans, University of California Santa Barbara, written communication, 2021) for rhyolite in the Grass Hills area indicate that they are about 21.5 to 22.0 Ma. The composition and orientation of these outcrops – particularly the long chain that cuts through the middle of the wilderness area inferred to be intrusive in part – also bears some similarities to the Eagle Crags Dike Swarm about 25 km to the northeast. There Sabin (1994) described a swarm of west-trending high-silica rhyolitic feeder dikes with up to 30% phenocrysts of quartz, sanidine, and plagioclase, plus biotite and Fe-Ti oxides. The swarm is up to 990 m wide and has a strike of ~281° (Andrew et al., 2014). Sabin (1994) dated one of the dikes, which yielded a 40Ar/39Ar age of 18.37 ± 0.06 Ma from sanidine. Determining the age of the Grass Valley rhyolites would allow comparison with ages of the Opal Mountain volcanics and Eagle Crags dike swarm and inform on the stress regime during emplacement.

North of the Grass Valley Wilderness Area, several north-trending, dark-weathering table mesas form the Black Hills. The mesas preserve units that range in age from about 19.6 to 7.2 Ma and thus record the transition from extension through the inception of the eastern California shear zone.

The southernmost mesa is a ~5 km long, up to ~2.5 km wide, and composed of a series of tabular dacite flows with acicular hornblende in a fine-grained matrix that varies from devitrified to glassy. Glass from the flow was dated by Osokin and Iriondo (2004) as late Miocene with a 40Ar/39Ar age of 7.23 ± 1.07 Ma. The dacite is ~100 m thick and generally well-foliated. It forms a low-relief mesa with arcuate ridges inferred to be pressure ridges. These observations suggest that the dacite erupted from a vent near the south end of the mesa and flowed north across a pediment of Mesozoic plutonic rocks overlain at least locally by volcaniclastic alluvium deposits (Felger, 2020). A conglomerate composed primarily of rhyolite boulders up to 4 m in diameter, and lesser, smaller boulders and cobbles of fine- and coarse-grained plutonic rocks is locally exposed under the west side of the flow. Pumice tuff and vitric lapilli tuff are also exposed under the west side of the flow. The conglomerate and tuff sequence is up to 60 m thick in places; however, its lateral extent is uncertain. Rhyolite boulders up to 4 m in diameter
have been observed in the talus apron on the east side of the mesa, suggesting that the conglomerate is laterally continuous under much of the dacite. The dacite flow and underlying pumice tuff have similar SiO$_2$ values of 64% and 66%, respectively. The other major oxides, trace, and rare earth element values also show similarities that suggest the dacite flow and the underlying pumice tuff are genetically related.

The mesas to the north of the dacite flow are, in aggregate, ~7.5 km long, up to 4.5 km wide, and capped by basalt with a $^{40}$Ar/$^{39}$Ar age of 11.66 ± 0.06 Ma from groundmass (Andrew et al., 2014). The basalt overlies basaltic andesite with a $^{40}$Ar/$^{39}$Ar age of 19.63 ± 0.32 Ma from plagioclase (Andrew et al., 2014), which in turn overlies either bedded ash and lapilli tuff, or Mesozoic plutonic rocks. The basaltic andesite-tuff sequence, which is correlated with the Eagle Crags volcanics to the east (Andrew et al., 2014), is laterally variable, so in places, the basalt overlies tuff and/or Mesozoic plutonic rocks. The high point of the mesa appears to be the vent area for the basalt. A dark, narrow ridge (visible on areal imagery) trends almost due south from the inferred vent across underlying rocks, and may be remnants of a feeder dike. If so, it may inform on the tectonic stress regime that existed at the time of eruption.

Northwest of the Grass Valley area lies the Lava Mountains and nearby hills, which are underlain by an extensive extrusive system ranging in age from 20 to 7 Ma (Smith et al., 2002; Andrew et al., 2014). The volcanic rocks younger than 19 Ma include the basalt and dacite described above from the Black Hills and similar flows in the Lava Mountains, together ranging from 11 to 13 Ma. Overlying rocks are dacite domes and lavas, lapilli tuff, and felsite intrusions dated at ~10 to 11 Ma. These in turn are overlain by the Almond Mountain Volcanics, composed of dacite domes and tuff about 7.5-8 Ma. The uppermost volcanic sequence is the Lava Mountains Dacite (which includes andesite), about 6.5 to 7.5 Ma. Distal equivalents of several Lava Mountains units are present in the Dove Spring Formation north of the Garlock fault (Smith et al., 2002; Andrew et al., 2014).

Dacite domes and plugs also are scattered in the western Mojave Desert (Fig. 1), where they form steep-sided buttes in a low-relief pediment-dominated landscape. Dibblee (1958) described dacite as pinkish-brown or gray, phenocryst-rich, plagioclase-dominated rock with lesser phenocrysts of sanidine and quartz. We consider that the presence of sanidine and quartz indicates a rhyolite composition. In addition, much quartz is black, suggesting radiation damage and in turn, U-rich rock. The domes form a linear array southeast of California City and are otherwise isolated. No age information is available. In the Boron area, buttes with steep to moderate slopes are widespread and underlain in many cases by the Saddleback Butte Basalt, which at ~20 Ma (Gans, 2022), is older than our time of interest.

The Red Buttes Basalt, an andesite farther south (Dibblee, 1960b) is described similarly, and is also ~21 Ma.

In the Rosamond-Soledad area of the western Mojave Desert (Fig. 1), several small (0.1-0.2 km$^2$) biotite-hornblende rhyolite domes and stubby flows lie within the Gem Hill Formation of the Tropico Group (Dibblee, 1963). These domes are undated and lie in a sequence of coarse-grained white tuffs and coarse pumice beds suggestive of local derivation. Farther north, several compound domes, also rhyolite, but strongly silicified, have been extensively mined for gold (Wilkerson, 2020). The compound rhyolite dome at Soledad Mountain (24 km$^2$) erupted between 21.5 ± 0.3 Ma and 16.9 ± 0.3 Ma; alunite in Middle Mountain to the west yielded an age of 18.36 ± 0.55 Ma (Blaske et al., 1991).

**Summary of rock chemistry**

Our limited geochemical data are strongly influenced by the data in the Lava Mountains of Smith et al. (2002) and data for the Yermo volcanic field (Singleton and Gans, 2008). Total-alkali vs. silica plots (Fig. 3A) show the rhyolites at Alvord Mountain and Grass Valley to be distinct with respect to total alkalis. However, the Alvord Mountain samples are reworked tuffs and may not represent magma chemistry completely. Overall, dacites range from 63 to 70 wt. % SiO$_2$ and 5 to 7 wt. % total alkalis. Dacites at Yermo, Cronese, and those reported by Glazner (1990) are similar to early Miocene dacite of the Eagle Crags field. The dacite at the Lava Mountains is slightly lower in SiO$_2$ and alkali values. Dacite of the Black Hills appears to be similar to that of the Lava Mountains. As previously noted, four dacite samples from the Yermo field are potassically altered, plotting as trachytic dacite. Smith et al. (2002) described andesite in the Lava Mountains, unique in our compilation at 59-60 wt % SiO$_2$. Besides the few andesite samples, there is a large compositional gap between basalts and dacites. Basalt in the Lava Mountains is similar to Pliocene basalt of the Black Mountain, Bicycle Lake, and Broadwell Mesa fields (Miller and Buesch, 2022). Basalt at Alvord Mountain differs with its lower SiO$_2$ values (~48 wt %).

The rocks have calcic to calc-alkalic affinities (Fig. 3B) except for one altered dacite from the Yermo field and biotite-rich rhyolite from the northern Grass Valley area that possibly has zeolitic alteration. The main trend in the data is like the trends from southern California and Sierra Nevada batholiths (Frost et al., 2001). Trace-element discrimination (Pearce et al., 1984) plots (Fig. 3C) indicate affinities with magmatic rocks in arc settings, with one rhyolite lying in the continental interior ‘within plate’ field. These plots indicate differences in dacites, such as the Cronese area domes, which are low in both Nb and Y, and the Yermo domes, which are high in Y. The designation of arc setting does not describe the tectonic setting of volcanism inboard of the San Andreas transform fault and probably reflects melting of crust produced during earlier subduction.
Summary of temporal patterns

Volcanic rocks in the area and time range of interest for this compilation have been dated by various methods. The dates are summarized in Figure 8, which shows dates binned in 2 m.y. intervals and distinguished by rock composition. Bearing in mind that the dating is heavily concentrated in the Lava Mountains and near Barstow, the 66 dates show that tuffs are common from 18.8 Ma to about 12 Ma. These may be local or regional in origin and do not necessarily reflect local volcanism. Dacite is the dominant composition and occurs in pulses from 20-16 Ma, 12-10 Ma, and 8-6 Ma, primarily reflecting the timing in two dacitic volcanic fields, the Barstow-Calico area for the older pulse and the Lava Mountains for the younger pulses. Dating dacites in the Cronese area and the western Mojave Desert is needed to better understand the temporal span of dacitic volcanism. Occurrence of andesite is rare and rhyolite lavas are uncommon and poorly dated. Until rhyolites in Grass Valley, Barstow, Soda Mountains, and the Rosamond areas are better dated, it is not possible to establish their contemporaneity with other silicic volcanism.

The post-Peach Spring Tuff (PST) volcanic rocks that are firmly dated demonstrate a long-lived field in the Lava Mountains from 12 to 7 Ma. The field is compositionally expanded from basalt through andesite and dacite to rhyolite. Other areas with numerous domes and flows are limited to dacite and rhyolite, based on field study and limited geochemistry. The ~17-Ma Yermo field in the Calico Mountains is about two m.y. younger than the Pickhandle Formation volcanic rocks that predate the PST; the field consists entirely of dacite, erupted over a brief 300,000 years. Nearby domes of the Mitchel Range (Fig. 2) and Barstow area are incompletely dated but include dates of both Pickhandle and Yermo ages. These domes are mainly dacite but range to rhyolite locally. Scattered domes on the Coolgardie Camp plateau are poorly dated but possibly range from Pickhandle age to as young as 13 Ma. Lastly, the Woods Mountains volcanic center is about 17.7-17.8 Ma. Elsewhere, groups of dacite and rhyolite domes are not well dated.

Summary of spatial patterns

The post-PST and possible post-PST rocks we have summarized form patterns that vary from south to north. At the latitude of Barstow, they are widely distributed, although with a few dense clusters, from the Woods Mountains on the east to Rosamond on the west. The Yermo field is the densest accumulation, but the Barstow and Coolgardie Camp plateau areas have several domes and flows each (Fig. 1, 2). Farther north, clusters of domes and flows, some intrusive, occur from Grass Valley north to the Lava Mountains. This accumulation apparently has no counterparts to the east or west. Although several groups of dated volcanic rocks coincide with older volcanic fields, such as in the Barstow area, Alvord-Cronese area, and Grass Valley, many other domes and flows erupted on bedrock and have no apparent relation to prior volcanism (Woods Mountains, Coolgardie Camp plateau, northern Grass Valley). In addition, there are expanses of early Miocene volcanic-dominated sequences such as the Pickhandle Formation that have no subsequent volcanic rocks (e.g., Newberry Mountains, Bristol Mountains). More dating of the Eagle Crags field is needed to determine if any late domes and flows are significantly younger than the bulk of the field. At this time, it is difficult to unambiguously determine the possible relation of young silicic volcanism to prior volcanism.

We have limited our investigation to regions west of the Colorado River corridor and its northward migration of volcanism. As eruption ages for the domes and flows we compiled are determined, the comparison with the Colorado River corridor will be important. Are the Mojave Desert volcanic rocks a western fringe of the Colorado River corridor volcanism? How do they relate to development of the San Andreas plate boundary farther west?

Summary and discussion

Dacite and rhyolite edifices that we compiled and are known to be or possibly younger than 18.8 Ma are present across much of the northwestern Mojave Desert but, except for the Woods Mountains caldera system, are absent from the area east of Baker and are apparently limited or absent in areas farther south in the Mojave. Eruption styles are dominantly effusive, producing isolated domes and flows and small dome fields. Most are dacitic, with subordinate rhyolite. This style and composition contrasts with previous early Miocene
compositionally expanded explosive and effusive volcanism. The dense cluster of dacite domes at 17 Ma in the Calico Mountains is unique although similar young dome clusters may exist at the Lava Mountains. Clusters of rhyolite domes and flows in the Rosamond area are poorly dated. Elsewhere, most edifices are isolated, and few are composite. Ages range from 17 Ma at Calico to 7 Ma in the northern Lava Mountains and Grass Valley, where rhyolite also is common. Most are undated.

The most likely origin for the silicic rocks we have highlighted in this paper is partial melting of crust, an origin that requires a heat source such as mafic melts generated from the mantle. The lack of basalts in many of the silicic eruptive complexes is notable. Future investigations that better establish age trends and improved understanding of the geochemistry of the silicic rocks may point to more widespread basaltic intrusion than is generally considered for the Mojave Desert. For instance, the Yermo volcanic field of dacite domes, at ~17 Ma, requires a magma parent at a time that few basalts are known. The transition from basalt- and andesite-dominated sequences pre-Peach Spring Tuff, and during at least localized crustal extension, to more silicic rocks may result from the inability of basalts to buoyantly rise through the entire crust after extension ceased (e.g., Mordick and Glazner, 2006; Loucks, 2021). By inference, basalt generation continued well after extension ceased.

Rocks sampled thus far present a chemical pattern of largely dacitic rocks, with SiO2 values of 63 to 69 wt %, but some range to high silica rhyolite. However, the dacites are distinct from a small group of andesites, which are distinct from the few late Miocene basalts. These patterns may indicate that mafic magmas rarely reach the surface but produce large amounts of dacite, spread over wide areas, through partial melting. Limited isotopic and geochemical study hampers petrogenetic interpretations, but studies thus far indicate crustal contamination of basalts as well as more silicic rocks (Glazner, 1990).

The resumption of basalt eruption ~6 Ma in a few widespread locations of the Mojave Desert (Miller and Buesch, 2022) may owe to a tectonic transition that promoted diking that allowed magmas to transect the lithosphere more readily. It will be important to establish contemporaneity or lack thereof for young silicic and basaltic volcanism, and to tie age patterns with location patterns for hints of tectonic origins.

Several tectonic transitions are known in the Mojave Desert area we studied, the most important of which are the end of localized extension in several places ~19 Ma and the transition to strike-slip tectonics of the early eastern California shear zone ~11 Ma. Other transitions are known from the study of plate tectonic data. A goal of our study, to be able to relate the volcanism to the macrotectonic regime, is challenging to accomplish with the paucity of dated volcanic edifices. It appears that compositionally expanded (basalt to rhyolite) magmatism accompanied the early “Barstow” timeframe of 19-16 Ma, after which dacite exceeded rhyolite for a silicic phase of volcanism that may have lasted to ~7 Ma but many domes and flows, especially in the Cronese and Rosamond areas, are undated. It is possible that silicic volcanism largely ceased except near the Garlock fault at the start of strike-slip tectonics ~11 Ma.

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Physical and geochemical architecture and age of the Pliocene Bicycle Lake basalt, southeastern Fort Irwin, California

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ABSTRACT—The informal Bicycle Lake basalt forms a volcanic field in southeastern Fort Irwin, California, disrupted by three east-striking faults and linked cross-faults of the Eastern California Shear Zone, and its distribution provides a framework for assessing volcanic field development, groundwater resources, and fault offsets. Previous geochronologic studies yielded ages ranging from ~2.9–5.6 Ma, and a new cycle of dating reported herein yields a more robust age of ~4.55 Ma. No vents or dikes have been identified for the basalt. No interstratified sedimentary rocks have been identified, not even aeolianites, implying that the field was monogenetic and short-lived. Thickness of the field varies from 1 m at distal edges to as much as 20–35 m (projected in cross-section) in the central part of the field. The field consists of lava flows <8 m thick, and the relative abundance of flows <2 m thick and 3–8 m thick varies across the field, as does the interstratification of <3 m thick cosets of <1 m thick flows. Most flows are basaltic andesite with minor basalt, all with olivine, plagioclase, and minor pyroxene phenocrysts. Within stratigraphic sequences, variations in major oxides and trace element geochemistry indicate interstratification of basalt with basaltic andesite, but also different compositions of interstratified basaltic andesite. The local abundance of basalt or different compositions of basaltic andesite varies. The distribution of compositions suggests minor changes in the composition of erupted magma and minor sector partitioning of flow paths during growth of the field. The basalt has been deformed as a result of transpressional stress transferred from the Coyote Lake to Bicycle Lake faults, which formed NW-striking reverse faults and faulted kink folds along with broad arching of the basalt. Restoring separation along the Coyote Canyon and Bicycle Lake faults indicates some flows of the basalt field traveled at least 10 km from their inferred source.

Introduction
The informal Pliocene Bicycle Lake basalt in southeastern Fort Irwin, California (Fig. 1), is being studied to understand (1) the physical and geochemical architectural development of the field, (2) the geologic and hydrogeologic framework for Fort Irwin, and (3) deformation in the Eastern California Shear Zone (ECSZ). Understanding the development of the basalt volcanic field enables comparisons of different parts of the field to determine directions of lava flow, potential runout length of lava flows, and the spatial and temporal growth and expansion of the field.

The basalt is exposed on four ridges, including (1) small fault blocks along the Coyote Canyon fault, (2) the ~0.5 x 3 km, slightly arched Bicycle Lake mesa (BLM) south of Bicycle Lake, (3) the ~2.0 x 7.0 km, southwest dipping dip-slope ridge west of Bitter Springs (BSW), and (4) the 1.7 x 4.2 km, sub-horizontal ridge south of Bitter Springs (BSS). The basalt is also present in boreholes CRTH1 and CRTH2 (Fig. 1, 2). There is no formal geographic name for BSW, but at Fort Irwin it is informally referred to as the Whale, and the name “the whale” was used for the ridge in the mid-1800s along the Old Spanish Trail (Lyman, 2004). There also is no formal geographic name for BSS, but it is herein informally referred to as the Squid because it is consistent with the aquatic theme, and it looks like one in aerial photos. Both the Whale and Squid are large areas that are herein divided into informal geographic domains, within which there are dip subdomains displaying characteristic dip angles (Fig. 2). Geologic mapping at various scales in the areas where the basalt crops out includes Byers (1960), Schermer and others (1996), Miller and Yount (2002), Miller and others (2014), and our recent mapping focused on the Whale, Squid, and BLM. The basalt as described by these workers consists of olivine, plagioclase, and minor pyroxene phenocrysts set in groundmass of varying grain size.

The Bicycle Lake volcanic field has been disrupted by faults of the ECSZ (Schermer and others, 1996; Miller and Yount, 2002; Miller and others, 2014) (fig. 1). Coyote Lake fault has an unknown amount of total separation (Miller and Yount, 2002), and at least one strand of the fault zone separates the Whale and the Squid. Bicycle Lake fault is traced from north of the Whale to near BLM where it bifurcates into northern and southern strands (Byers, 1960; Schermer and others, 1996; Miller and Yount, 2002),...
and the northern strand might transition into northwest striking faults that project to the Coyote Canyon fault. In detail, Schermer and others (1996) mapped four strands numbered from 1 in the south to 4 in the north. Strands 1–3 combined have >4.7 km of left-slip separation on a band of Proterozoic (?) marble (also reported by Byers, 1960), and all four strands combined have >6 km of left-slip separation on the western border of Jurassic granite. The structural block containing the BLM and basalt is bounded by strands 1–3, and the BLM is 12–16 km west of the Whale. Schermer and others (1996) proposed an apparent separation of 3–8 km for the basalt relative to the Whale, and the distances greater than the apparent separation represent minimum distances the basalt flowed. Coyote Canyon fault has an estimated 4.1 km of left-lateral separation, and the basalt is between the two northern strands of the fault (Schermer and others, 1996).

The age of the basalt has been challenging to determine. Schermer and others (1996) reported a 40Ar–39Ar age of 5.6 Ma (an average of two dates of 5.57 ± 0.26 Ma and 5.50 ± 0.20 Ma from the Bicycle Lake mesa and Whale, respectively). Miller and Yount (2002) published an average 3.4 ± 0.5 Ma K-Ar age from the Coyote Ridge area (just north of the Coyote Canyon fault). There was an unpublished K-Ar date of 2.9 ± 0.2 Ma (J.K. Nakata, insert agency name/affiliation, written communication, 1995 in Miller and Buesch, 2022 this volume) also from the Coyote Ridge area. Additionally, a fine-grained ash above the basalt flow has a 3.4 ± 0.2 Ma K-Ar date on biotite and a tephrochronologic age of 3.4 Ma based on glass chemistry correlations (Miller and Yount, 2002). Recently, we determined a 40Ar/39Ar eruption age of 4.55 ± 0.07 Ma on a basalt flow from the southwestern part of the Whale, as described in detail later.

**Physical stratigraphy**

Along the south and northeast parts of the Squid and the east side of the Whale (fig. 2), the basalt overlies fine-grained sedimentary deposits representing playa, groundwater discharge, and distal alluvial fan environments. These include wetland deposits south of the Squid that have preliminary tephrochronology ages.
of between 9.3 Ma to 6.3 Ma (Walkup and others, 2022). The implication is that the basalt (on Whale and Squid ridges) was deposited on very low-relief and shallowly sloping surfaces. Recent mapping at the Squid and northeast Whale indicates that locally there is sandstone and conglomerate up to 2 m thick beneath the basalt that might indicate a period of incision and channel filling between the deposition of the fine-grained deposits and the basalt. At BLM, the basalt was deposited on Jurassic (?) plutonic and metamorphic rocks (Miller and others, 2014), and locally, on relatively thin sandstone and conglomerate with polymictic (granitoid and metamorphic) clasts. Southeast of BLM are several small exposures of the basalt in fault blocks along the Bicycle Lake fault, and pre-basalt conglomerate in the Tiefort Mountains is composed of clasts sourced from granitoid and metamorphic rocks (Miller and Yount, 2002). Throughout the entire basalt field, there is no evidence of any sedimentary rocks (even aeolianite) interstratified with the lava flows. The apparent lack of any clastic sedimentary incursions into the volcanic field as it developed implies a relatively short-lived field.

Because of the distribution of exposures, the thickness throughout the basalt field is difficult to determine; however, the thickest sections are exposed at the Whale, with thinner sections at the Squid and BLM, and thin remnants at other places. Lava flows at the Whale are well exposed, but most are exposed along dip slopes that have the uppermost chilled tops (such as those with ropy surfaces on pahoehoe) eroded or have desert pavements. The best exposures are along the edges of the Whale or in minor canyons that typically are incised <10 m into about 18 m in the basalt. At the southern Whale, the best exposures are in minor canyons with a maximum incision of <16 m, and projections in cross-sections through the southern Whale indicate thickness might be 20–35 m thick. The basalt in borehole CRTH1 is ~12 m thick at a depth of 73 m interpreted from gamma and resistivity logs of Kjos et al. (2014). Along a probable margin of the field in the Squid and northeast Whale, each has a single ~1.5 m thick flow. In the northern Whale, the maximum incision into the basalt is <18 m, and at the northeast edge of the Whale (northeast of a 285° striking fault) the basalt section dips 45–90° and is 1–10 m thick. In borehole CRTH2, which is north of, or possibly within, the Bicycle Lake fault zone, the basalt is at least 24 m thick at a depth of 293 m based on interpreted gamma and resistivity logs of Kjos et al. (2014). In the BLM area, where the basalt flow filled a 300–600 m wide paleochannel incised into Jurassic rocks, the basalt varies from 0.5 m thick on the paleochannel margins and to 20 m thick in the center. The paleochannel bottom is locally exposed in gullies eroded through the basalt, where sandstone and conglomerate are exposed beneath the basalt.

The textures, structures, zoning, thickness, and morphology of lava flows provide insights into how the lava flowed. Lava flows, many with flow tops having rropy pahoehoe textures, typically are vertically zoned based on vesiculation, groundmass grain size, and fracture characteristics (Fig. 3). Vesicles vary in size and abundance, and the geometry varies from spherical to oblate or prolate, and some indicate the direction of flow. Groundmass variations include glassy and very

Fig. 2. Map of geographic domains and dip subdomains on ridges Whale and Squid, and Bicycle Lake basalt sample locations and compositions. Sample locations are indicated at symbols; however, where closely spaced or duplicated, the location is indicated by a black dot and the composition is indicated by the adjacent balloons. Geographic domains (on one or both Whale and Squid): central east, CE, central west, CW; north central, NC; northeast, NE; northwest, NW; southeast, SE; southwest, SW. Dip subdomains: 1, 2, 3, 4, and 5. Inserts show radar plots (as in Fig.1) for lava flow directions in the north Whale, southwest Whale, and Squid. Base image is from a Digital National Agriculture Imagery Program (NAIP)† image with 1 m (pixel) ground resolution. The image is displayed in ArcGIS®, and the map image is the intellectual property of Esri and is used herein under license. Copyright © 2020 Esri and its licensors. All rights reserved.

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fine-grained (chilled subzones) at the top and bottom of the flow with coarser groundmass in the core. Cooling fracture geometry and surface planarity and roughness form subzones. For example, fractures that parallel the top and bottom of a flow and those that are perpendicular to the base are closely spaced at the top and bottom and are more widely spaced in the core. Also, the fracture surface planarity (shape at the decimeter or greater lengths) and roughness (shape at the millimeter and lesser lengths) near the top and bottom of flows differ from those in the center. At both the southern and northern Whale, many lava flows where measured in the field are 1–2 m thick and a few are locally 3–8 m thick, and these are interstratified with 0.5–3.0 m thick cosets of 0.1–1.0 m thick individual flows. [Paraphrasing from Fisher and Schmincke (1984) and Jackson (1997), a co-set (or coset) is a set or sequence of beds that have similar characteristics such as texture, structure, or composition that set them apart from other beds above or below. Although defined for sedimentary beds, this architectural descriptor can be applied to sequences of lava flows.] The southern Whale has slightly thinner flows than the northern Whale with the typical maximum thickness of 6 m, and more <3.0 m thick co-sets (Fig. 4); however, one flow (1–2 m thick) increases to 10–12 m thick within ~5 m along strike (a highly inflated flow). In contrast, the northern Whale has slightly thicker flows with the maximum thickness of 8 m including several 3–5 m thick flows that have rounded but moderately steep margins (inflated pahoehoe flows), and a few localized <3.0 m thick cosets exist. No lava tubes have been identified, and some flows developed tumuli. Locally, granitoid xenoliths occur in the basal flows (probably scoured clasts from substrate conglomerate?), and they are also in one flow high in the section in the northern Whale. Flow morphology implies low-viscosity (high-temperature?) lava that filled low-relief topography developed on previous flows.

Based on individual flows where lengths are limited by truncation at a fault, lack of exposure, or termination of a flow (possibly influenced by topography of a preexisting flow), individual flows are low relief and traceable for hundreds of meters where measured in the field, and they have a moderate to high aspect ratio (length/thickness) of 40 to 125. However, using the basalt exposed on the Squid, Whale, BLM, and Coyote Ridge (Schermer et al., 1996; Miller and Yount, 2002; Miller et al., 2014), where field measured stratigraphic sections (1–20 m thick) are examined relative to the length of exposure between locations (the inference being an aggregate of flows needed to travel this distance), they have high to very high aspect ratios (length/thickness). At the Squid aspect ratios are 290–4,300, and at the Whale they are 375–7,500. Where the possible separation along the Bicycle Lake fault has been removed, the aspect ratios from the Whale to BLM are 650–4,000, and Whale to Coyote Ridge are 4,000–10,000.

Direction of flow of lava deposits can be measured or inferred from field relations of large features (tens to hundreds of meters of exposure) such as orientations of channels and levees or “ropy” and “ridge” structures on the tops of lava flows, or small (mm to dm) structures including imbrication of vesicles relative to the base of the flow and prolate (vertically elongated) vesicles bent by the lava moving past the congealed base of the flow. In the Bicycle Lake basalt, channels and levees are not easily identified or separated from a local onlap of one flow against another. The lava flows were pahoehoe, and many formed ropy structures that have been used for flow direction indicators. Imbrication of oblate vesicles relative to the base of the flow form where vesicles are deformed (elongated) by shear induced in the lava from the velocity gradient between the zero-velocity congealed lava base and the moving inner part of the lava. The imbrication of oblate vesicles creates a planar fabric that can yield
a vector direction of flow. Many lava flows developed a lower prolate vesicle zone (up to ~20 cm thick) just above the chilled base and in some the prolate vesicles continue upward into the overlying sparsely vesicular core. Prolate vesicles develop a vertical profile where the vesicle is ~90° to the base but upwardly can curve (bend) to angles less than 40° to the base. Measuring the direction and plunge of the bent prolate vesicle indicates the direction of flow that formed that vesicle. Measurements can be combined from differently oriented fracture surfaces to enable calculation of the flow vector. Each directional flow indicator at a site provides local detail about the flow; however, lava flow morphology can be complicated as small lobes or breakouts develop (possibly changing or redirecting the direction that the lava originally moved), new distributaries form, or where flow was around an edge of a previous lava flow. We combine several measurements, when possible, to better interpret overall flow direction.

Measuring flow direction indicators in different parts of the field, and vertically in lava sequences, can provide a sense of how and where the lava flowed, and conversely where it might have flowed from. Based on field measurements and photographs, 46 locations from five geographic domains are summarized (including radar graphs in Figs. 1 and 2) (Buesch, 2022). Flow directions were grouped in 22.5° bins. In BLM, the lava flows were along the paleochannel to the NNW-N, with one example of channel overspill to the ENE. This location appears to be part of a paleochannel trending to the N-NNW that changed orientation from a long WNW-trending paleochannel (Fig. 1). In the north Whale, most flow directions in ncWa1 were to the SSE-S with scatter in several directions WSW-ENE, and in neWa4 most flows were to the ENE-ESE with one flow to the SSW (Fig. 2). These two subdomains are geographically separate, and there is a fault with unknown separation along the southwest side of neWa4. In the south Whale, seWa1 had no directional information, swWa2 had flows to the WSW and NE, swWa3 had most flow directions to the SE (with a range ESE-SSW) and a few directions NE-ENE and NNW, swWa4 had one measured flow to the NNE, and swWa5 flowed to the SE (Fig. 2). In the Squid, most flows were to the SSE-SSW with a few WSW-WNW. The Whale and Squid were both in the southern part of the field, and the Squid was at the distal end of the field. Because flows in the southern Whale were to the ESE-SSW with most SE, and most flows in the Squid were to the SSE-SW, it is possible that flows that went through or near to the southern Whale might have continued but were redirected to the SSE-SSW by paleotopography. Alternatively, the Whale and Squid might represent different parts of the southern basalt field that have been juxtaposed by the Coyote Lake fault.

The eruptive source of the Bicycle Lake basalt appears to be unexposed. Our mapping and aerial photo interpretations have not identified any scoria cone or effusive vents. However, a simplified summary of basalt flow indicators suggests where the eruptive center might have been. This simplification does not overly weight the apparent locally redirected flows, and the directions have not been corrected for possible separation along faults such as the Coyote Lake fault. Flows in the Squid came from the NNW-NE, but these directions might be local paleographic control at the distal ends of the field. Flows in the southern Whale came from the NNW-WNW, flows in the neWa4 came from the W-WSW and represent less than five flows in the sequence, flows in the ncWa1 came from the N-NNW (but with substantial variation), flows in BLM came from the ESE along a long paleochannel that changed orientation to the NNW-N, and flows in Coyote Ridge are presumably the distal end of the paleochannel at BLM. These paleoflow directions are very generalized, and more mapping is needed in the northwest Whale, but projections suggest an eruptive center somewhere W-NNW of the northern Whale and possibly offset by the Bicycle Lake fault. In either case, an eruptive center appears to be buried by valley-fill sediments. Schermer et al. (1996) also concluded that the Whale area was the probable source for lava based on its thickness, and we support and refine this interpretation by using the paleoflow directions and borehole thicknesses that are similar to or exceed those at the Whale. Together, the boreholes and outcrop indicate a much greater volume of lava in the Whale vicinity.
Geochemical composition

As displayed on a Total Alkali Silica (TAS) diagram (Le Maitre, 1989), Figure 5 shows that the Bicycle Lake basalt is mostly basaltic andesite with a few flows of high-SiO$_2$ basalt (Buesch, 2022). This diagram also shows that the composition of the Bicycle Lake basalt differs from that of most other Miocene and Pliocene basalts and basaltic andesites in the Mojave Desert area. Only one sample of the Bicycle Lake basalt overlaps with the trend of Miocene basalts and basaltic andesites from western Fort Irwin. The two analyses from the Miocene Eagle Crags volcanic field (Sabin, 1994) are along the fringe of the Bicycle Lake basalt compositions. There is no overlap of Bicycle Lake basalt with the Miocene basalt in the Barstow Formation, the Marble Mountains, or the 5.4 Ma Broadwell Mesa basalt (Buesch, 2017). The ~3.9 Ma Black Mountain basalt (Miller and Buesch, this volume) is represented in Fig. 5 by samples from Black Mountain and from the Hinkley and Water Valley areas. Compared to the Bicycle Lake basalt, the Black Mountain basalt has a similar range in SiO$_2$ wt% with a slightly higher Na$_2$O+K$_2$O wt%, and there is ~15 percent overlap of these two fields.

Bivariate plots of major and minor oxide concentrations, trace elements, and elemental ratios show the basalt has compositional variations locally and throughout the field (Buesch, 2022; Figs. 6–8). To minimize any effects of alteration, the oxide and trace element concentrations have been mass normalized using the methods of Sawlan (2018). TiO$_2$ and Al$_2$O$_3$ are effectively immobile elements, so they (and the Al$_2$O$_3$/TiO$_2$ ratio) are used as a common reference on the y-axis, and the x-axis displays MgO, Cr, and Zr as examples of minor oxides and trace elements (Figs. 6–8). These oxides, trace elements, and oxide and element ratios have been used to demonstrate compositional variations in the Broadwell Mesa basalt (Buesch, 2017) and the Columbia River basalt (Sawlan, 2018). Repeat analyses on splits from five hand specimens (using six pairs) are used to approximate the analytical uncertainty of our geochemical analyses (Figs. 6–8). The average differences (uncertainty) $\pm$ 10 in analytical uncertainty for TiO$_2$ is 0.04 $\pm$ 0.02 wt%, Al$_2$O$_3$ is 0.05 $\pm$ 0.03 wt%, MgO is 0.10 $\pm$ 0.09 wt%, Cr is 11.73 $\pm$ 5.34 ppm, and Zr is 10.76 $\pm$ 9.18 ppm. Variations in the composition of lava flows within local flow sequences and variations between different regions of the basalt are larger than analytical uncertainties based on replicate analyses, which indicates that there are small but real variations in the composition within the basalt.

The Al$_2$O$_3$/TiO$_2$ ratios (A/T) have three categories; low ratio (LR) is 5.1–6.0, typical ratio (TR) is 6.0–6.6, and high ratio (HR) is 6.6–7.2 (Fig. 9). Basaltic andesite typically has TR with some HR, and basalt typically has LR. Basalt and basaltic andesite occur in different clusters on A/T plots; however, there is one "outlier" analysis of basalt in the basaltic andesite cluster, and one "outlier" basaltic andesite in the basalt cluster. These two "outlier" analyses are each spatially located within or adjacent to geographic domains in the Squid and northern Whale that are the other type of composition (basalt or basaltic andesite) and A/T ratios (Fig. 2 and 9), and therefore may be useful chemical marker flows. However, these two outlier samples plot close to the basalt and basaltic andesite boundary on

Fig. 6. Graph of bivariate plot of mass normalized TiO$_2$ mn versus MgO mn. Repeat analysis of sample splits have tie lines. Geographic domain and subdomain abbreviations are concatenations of the symbols listed in Fig. 2, except Sq for undivided Squid, and BH for borehole.
Fig. 7. Graph of bivariate plot of mass normalized TiO$_2$ versus Cr. Repeat analyses and domain symbols as in Fig. 6.

Fig. 8. Graph of bivariate plot of mass normalized Al$_2$O$_3$/TiO$_2$ versus Zr. Repeat analyses, and domain symbols as in Fig. 6.

Fig. 9. Graph of bivariate plot of mass normalized Al$_2$O$_3$/TiO$_2$ versus stratigraphic position in geographic domains. In each domain or subdomain, the relative position is the approximate percent of the exposed section with the lowest lava flow in the section on the left and the highest on the right. Typically, there are lava flows in the section that were not sampled. Samples are typically basaltic andesite (not labeled), and samples of basalt are labeled. Three categories of Al$_2$O$_3$/TiO$_2$ ratio; high ratio (HR), typical ratio (TR), and low ratio (LR). Repeat analyses; R1 was analyzed in the initial analytical run (R0, not labeled), and R2 was analyzed in a subsequent analytical run to R0.
top of the exposed section. The samples are basaltic andesite, and the lowest flow has the new $^{40}\text{Ar}/^{39}\text{Ar}$ date reported below. The second and third flows are indistinguishable in A/T and TiO$_2$, and the first flow is increased in A/T by 0.59 wt% (so, HR) and decreased in TiO$_2$ by 0.15 wt%. In MgO, the first and third flows are indistinguishable, and the second flow is increased by 0.51%. In Cr, the flows are indistinguishable. In Zr, the second and third flows are indistinguishable, and the first flow is decreased by 25 ppm.

- In Squid (Sq), in the NE domain, the first two samples are from the two lowest sequential lava flows. In the southwest part of the SE domain, there are only two lava flows (samples 4 and 5) that might represent the margin of the basalt field. The third sample is from the central part of the ridge, so its detailed stratigraphic position is not known. Four samples (1–4) are basaltic andesite, and sample 5 is basalt (one of the “outliers” described above). Samples 1, 2, and 5 are indistinguishable in A/T (TR), TiO$_2$, MgO, Cr, and Zr; however, sample 1 has lower MgO and Zr, and sample 5 has slightly higher MgO and Zr and slightly lower Cr. Sample 4 is an A/T TR, but has increased MgO by 0.75 wt%, Cr by 46 ppm, and Zr by 31 ppm. Sample 3 is an A/T HR (based on lower TiO$_2$ and higher Al$_2$O$_3$), but is otherwise indistinguishable from samples 1, 2, and 5 in TiO$_2$, MgO, Cr, and Zr.

- In Whale subdomain ncWa1, the base of the stratigraphic section is not exposed, and the top was not sampled; however, the nine samples represent a continuous section of lava flows that are each 2–6 m thick with thinner flows or flow cosets not sampled. Sample sequences 1–4 and 5–7 are the same (or similar) sections sampled along different traverses, and sequence 8–9 (with flow 9 near the top of the section) is ~250 m from the other sequences and across a possible fault. The sequence represents at least one cycle of lava flows with a stratigraphically upward change from basaltic andesite to basalt and back to basaltic andesite. Samples 1, 5, and 9 are basaltic andesite with A/T TR in the upper range of TR, and increase up section in TR, MgO, Cr, and decrease slightly in Zr. Samples 2–3-4 and 6–7–8 are basalt with A/T LT that decrease up section in TR, MgO and Cr, and increase in Zr.

- In Whale subdomain nwWa1, the single sample has no stratigraphic context (from the base of the topographic slope). The sample is a basaltic andesite with A/T LR and is one of the “outliers” described above. Regardless of plotting as basaltic andesite on a TAS diagram, the A/T LR, MgO, Cr, and Zr are more similar to the basalt samples from the nwWa1.

- In the boreholes, both samples are basaltic andesite with A/T TR and similar MgO, Cr, and Zr to other TRs. However, CRTH1 has low MgO and Cr, and CRTH2 has the lowest MgO (6.65 wt%) of any basaltic andesites.

- In domain BLM, the two samples are ~240 m apart and both are deposited on the local substrate, but it is not clear which was deposited first. Both samples are basaltic andesite. One sample has a A/T TR, and MgO and Cr are low, but it is similar to sample 1 in the subdomain ncWa1. The second sample has a A/T HR, MgO is a bit high compared to other HRs, Zr is the second lowest of all samples, and the A/T, TiO$_2$, Al$_2$O$_3$, MgO, Cr, and Zr are similar to sample 1 from subdomain swWa4.

To summarize, geochemistry from across the Bicycle Lake basalt field indicates that most of the lava flows are basaltic andesites that are locally interstratified with basalt. Most basaltic andesite with A/T TR and associated amounts of MgO, Cr, and Zr differ slightly from less abundant basaltic andesite with A/T HR and associated amounts of MgO, Cr, and Zr. In the south Whale (and in part of the Squid), currently available data (Buesch, 2022) indicate a general sequence (from bottom to top) of TR to HR to TR flows; however, additional interstratified flows have been collected, and the analyses have not been completed. In the north Whale, basalt with A/T LR and associated amounts of MgO, Cr, and Zr is interstratified with TR basaltic andesite. The sequences of basaltic andesite and basalt each have small up-section increases or decreases in A/T, MgO, Cr, and Zr that indicate changes in the erupting lava with time. In the BLM domain, about 12 km west of the north Whale, one TR basaltic andesite flow is similar to a flow at the north Whale, and a second HR basaltic andesite flow is similar to a flow at the southwest Whale. This is not to claim they are the same flows, but that similar flows can be identified across these distances.

Eruption age

Several studies have attempted to constrain the eruption age of the Bicycle Lake basalt, with conflicting results. Schermer and others (1996) reported $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 5.57 ± 0.26 Ma and 5.50 ± 0.20 Ma from the Bicycle Lake mesa and north Whale, respectively. Full analytical data corresponding to these $^{40}\text{Ar}/^{39}\text{Ar}$ ages have not been published and therefore the quality of these experiments cannot be assessed. Miller and Yount (2002) reported a K/Ar age of 3.4 ± 0.5 Ma for the basalt in the Coyote Ridge area north of the Coyote Canyon fault and a K/Ar age of 3.4 ± 0.2 Ma for a fine-grained ash found in alluvial gravel and paludal sediment stratigraphically above the basalt. The ca. 3.4 Ma age for the fine-grained ash agrees with tephrachronologic correlation based on glass chemistry that links it to a 3.4 Ma ash in Fish Lake Valley, Nevada (Miller and Yount, 2002). Thus, the K/Ar age of 3.4 ± 0.2 Ma on this fine-grained ash likely represents a robust minimum age for the basalt. An unpublished K-Ar age of 2.9 ± 0.2 Ma for the basalt (J.K. Nakata, USGS, written communication, 1995 in Miller and Buesch, 2022) is likely too young. The discrepancy between the eruption ages for the basalt reported in Schermer and others (1996) and
Miller and Yount (2002) cannot be explained based on the available data.

Recently, a sample of basalt from swWa4 (Fig. 2) was dated by the $^{40}$Ar/$^{39}$Ar method at the Argon Geochronology Lab in Menlo Park, Calif. The sample was crushed using a roller mill, ultrasonicated, and sieved to 250 to 355 μm. Groundmass was concentrated from the crushed sample using a Frantz magnetic separator to target a magnetic fraction that yielded feldspar-rich groundmass while excluding phenocrysts and magnetite-rich parts of the groundmass. The groundmass separate was then purified by handpicking under a binocular microscope. The sample was packaged in aluminum foil along with the Bodie Hills sanidine monitor mineral (9.7946 ± 0.0031 Ma, equivalent to Fish Canyon sanidine at 28.099 ± 0.013 Ma; Fleck et al., 2019) and encapsulated in a quartz vial along with other samples. The quartz vial was wrapped in 0.5 mm thick cadmium foil to shield samples from thermal neutrons and irradiated for 1 hour in the central thimble of the U.S. Geological Survey TRIGA reactor in Denver, CO (Dalrymple et al., 1981) at a power level of 1 MW. The sample vial was rotated and oscillated vertically around the center line during irradiation to minimize the flux gradient across the samples.

Following irradiation, monitor minerals were analyzed by laser total fusion using a CO$_2$ laser coupled with a MAP216 mass spectrometer to determine J values. The groundmass separate was rewrapped in degassed Ta foil and argon was extracted incrementally in eight discrete temperature steps using a diode laser coupled with a MAP216 mass spectrometer. Temperature was monitored using an optical pyrometer. Prior to measurement of Ar isotopic composition, the sample was degassed at 400°C until undesirable gases (e.g., water, nitrogen, and hydrocarbons as measured by a Granville-Phillips 835 VQM) were reduced to acceptable levels. The extracted Ar was exposed to a 4 A tungsten filament, a 125K cold finger, and two SAES ST-175 getters (one operated at 300°C and one at room temperature) to remove active gases prior to measurement.

The sample yielded well-defined plateau and isochron ages, which overlap at the 1σ level (Fig. 10). The initial trapped gas $^{40}$Ar/$^{36}$Ar age of 300.9 ± 4.6 (2σ) calculated from the isochron is within error of atmospheric argon (298.56 ± 0.31; Lee et al., 2006), and therefore we chose the plateau age (4.55 ± 0.07 Ma) instead of the isochron age (4.33 ± 0.17 Ma) for this sample. This new eruption age for basalt falls between the eruption ages reported in Schermer and others (1996) and Miller and Yount (2002). The discrepancy between this new age and those reported in the literature cannot be determined based on available data, but the fact that our new $^{40}$Ar/$^{36}$Ar age is based on 100% of the gas released during the experiment and contains trapped gas in equilibrium with atmospheric argon suggests these new results are robust. Further dating experiments on different localities of the basalt would be needed to determine the overall age range of this unit.

Structural deformation

The Bicycle Lake basalt is distributed across three east-striking left-slip faults of the ECSZ (from south to north); Coyote Lake, Bicycle Lake, and Coyote Canyon faults (Fig. 1). The Coyote Lake fault has several splays, two of which largely bound the Squid. There are some small faults in the Squid; however, the ridge overall describes an E-W elongate, 30–50 m relief, shallowly E-plunging
(gentle) anticline. The north side of the Squid is a slightly asymmetric N-vergent arch, and there is a central syncline represented by the embayment on the west end of the ridge. Less than 1 km north of the Squid is the southern end of the Whale, and there is a fundamental difference in the structural geometry across this gap.

The Whale is partitioned into three deformed regions: (1) southern, including domains SW and SE, (2) central, including domains CW and CE, and (3) northern, including domains NW, NC, and NE (Fig. 2). These domains and subdomains can be identified in the field and on lidar (Buesch, 2019; U.S. Army, 2009). The southern region overall forms a gently (4–12°) southwest dipping dip-slope ridge that is partitioned into dip subdomains 1–4, and basalt geometry in subdomain swWa5 and borehole CRTH1 are consistent with southwest dipping dip-slopes. The northern region (domains NW, NC, and NE) forms a 60–100 m relief, slightly E-W shortened arch, ~7° north-plunging fold. Along the west side of domain NW is a steeply dipping subdomain 4 (nwWa4), a continuation of swWa4. The northeastern region (subdomain neWa4) is bounded by an ~115° striking and nearly vertical fault with 1–4, thin (1–6 m thick) basalt flows and subjacent sedimentary rocks that dip 40–90° NNE. Inferred steeply dipping fault-bounded blocks and faults are needed to explain basalt to depths of 293 m in borehole CRTH2 and are consistent with down to the NNE offset along the Bicycle Lake fault. Alternatively, there may be an undetermined amount of strike-slip (or oblique-slip) separation along the Bicycle Lake fault or W-WNW-striking splays that brought these steeply dipping panels to this position. The central region (domains CW and CE), which includes the western dip subdomains cwWa2, 3, and 4 that are similar to those in domain SW, gradationally accommodates the transition in deformation from southern to northern regions.

Along the southwest side of the Whale, a curving set of kink folds have steep limbs that face SW and create a SW-vergent train of folds that are highlighted by light-gray narrow aeolian sand decorations (Fig. 2). As identified in the field and on lidar (U.S. Army, 2009), these folds are expressed as treads and risers in a stair-step topography that curves from trends of ~270° in the south to ~340° in the north. Most strain is accommodated by folds along the 25–90° SW-dipping subdomain 4 and 0–10° SW-dipping subdomain 3 that bound the western side of the ridge (domains SW, CW, NC, and NW) and is consistent with interpretations by Schermer et al. (1996) (Fig. 2). This structural doublet is interpreted as a faulted SW-verging kink fold. Although mapped in the field and based on the deformation of the basalt that forms erosionally resistant surfaces and landforms, the 1-m lidar hillshade and slope map shows the four subdomains (U.S. Army, 2009; Fig. 1). Both domains SW and SE are formed by the gently (4–12°) southwest dipping dip-slope of lava flows (or similar looking lava flows) that can be traced 300–500 m up slope. In addition to subdomains 4 and 3, at several locations up slope to the ridge crest, there are narrow bands of lava flows that dip 15–25° with adjacent 0–6° dipping lava flows, and the highest elevation set of 15–25° and adjacent 0–6° bands forms the boundary of subdomains 2 and 1. These bands are laterally continuous across the slope, and some are traceable more than 2 km. These slope breaks appear to be small versions of the larger faulted fold (subdomains 4 and 3) along the southwest margin of the ridge. This kink and faulted fold geometry is consistent with NE-SW compression that has been locally developed in the ECSZ (Schermer et al. 1996). Compared to the remainder of the basalt field, these fold trains indicate enhanced strain near the Coyote Lake fault. One possible interpretation is that the NW-striking faults forming the steeply dipping structural subdomain along the west side of the Whale are a transpressional transfer of strain that has buckled the Whale in a right step between the Coyote Lake and Bicycle Lake faults. Another is that the bending of the NW-striking fault, and the additional smaller kink and faulted folds, results from left lateral drag across the southern Whale along the Coyote Lake fault.

The Bicycle Lake fault extends west from the northern steep truncation of basalt at the Whale. At several positions the fault has multiple strands, between which small, tilted exposures of the basalt occur (Fig. 1). About 12–16 km WNW of the Whale, the fault zone steps northward via curving strands, between which lies the Bicycle Lake mesa with exposures of the basalt (Schermer et al. 1996; Miller and Yount 2002). Topographic profiles of the mesa top indicates that the flows have been deformed and arched, as suggested by Byers (1960). Topographic profiles from Google Earth images were drawn along the flow margin remnants of the underlying paleochannel exist. In the north, the flows and underlying channel have been arched as much as 30 m, and in the south the southernmost basalt is 92 m below the crest of the mesa, requiring arching of that amount plus additional deformation necessary to establish an original northward gradient. This arching might have resulted from transpression in the right step of the fault.

About 6–8 km northwest of Bicycle Lake mesa are fault-bounded exposures on Coyote Ridge and in the Coyote Canyon fault (Schermer et al., 1996; Miller and Yount, 2002). In the Bicycle Lake basin, there are numerous wells to various depths. Borehole-well BLA5 is 113 m deep, penetrates granitoid rocks in the lower part of the well, and although it is located along the projection of the basalt exposure from BLM, no basalt was penetrated as interpreted from gamma log values and the lack of low values that are indicative of basalt (Kjos and others, 2014; Buesch, 2018; Nawikis et al., 2019). Tracing faults through the Bicycle Lake basin is largely inferential, but northwest-striking faults occur south of, and splay into, the Coyote Canyon fault. Exposures of the basalt at Coyote Ridge, BLM, and Whale were mapped by Schermer et al. (1996; Miller and Yount, 2002), and topography from Google
Earth’s images, indicate that at Coyote Ridge, basalt is exposed at ~1,050 m, the highest elevation for any of the basalt. In comparison, the north end of BLM basalt is at 765 m, and at the Whale the highest elevation is ~650 m. Along this transect, ~70 percent of the elevation change occurs between Coyote Ridge and BLM. Considering that the basalt flowed approximately northwestern down a topographic slope across the area at Bicycle Lake mesa, even with a 1° slope, there would have been ~430 m of difference in elevation from the inferred source area near the Whale to the Coyote Ridge area. This requires a reversal of relief by more than ~830 m over ~25 km distance during the last 4.5 m.y., and this contributed to changes in drainages, paleoflow directions, and provenance for sediment described by Miller and Yount (2002).

**Summary and discussion**

The physical and geochemical stratigraphy of the Bicycle Lake basalt provides insights into the architecture and growth of this small volcanic field. Furthermore, as described by Buesch (2018), the basalt field influences our understanding of lava-flow hosted aquifers in groundwater resource assessment, especially because basalt lava flows can extend significant distances into basins (Buesch, 2018). At Fort Irwin, basalt lava flows are in several wells including CRTH1 and CRTH2, and Nawikas et al. (2019) summarized the hydrologic testing, well boreflow data, and water quality for 2011–2015 showing the rock and hydrogeologic unit with the highest hydraulic conductivity (K) and transmissivity (T) are the basalt lava flows. Studies of the Bicycle Lake basalt and how it has been disrupted by fault refines interpretations of fault systems in the ECSZ.

Lava flows occur as individual flows locally interstratified with lava flow cosets. Exposures of lava flows have low-relief and moderate to high aspect ratios of 40 to 125. Vesicle size and geometry indicate the ability for vesicles to form, migrate, and deform during flow. Chilled (fine-grained to partially glassy groundmass) tops and bottoms and well-developed fractures that parallel the flow top (and locally the base) indicate significant rates of cooling. All of these textures and structures indicate the lava flows were hot and/or low viscosity. One implication is that as the lava flow distribution system developed, the magma was capable of flowing long distances (possibly 20–25 km), even across what appears to have been shallowly sloping ground surfaces such as playa, distal alluvial...
fan environments, or to fill paleochannels leading to those places. Throughout the entire basalt field, there is no evidence of any sedimentary rocks (even aeolianite) interstratified with the basalt lava flows. This indicates no clastic sedimentary incursions into the volcanic field as it developed, implying a relatively short-lived field.

Using lava flow morphology and detailed flow direction, indicators of vesicle imbrication or bending enables reconstruction of the local direction the flow moved, and these can cautiously be back-projected indicating where the flow probably came from. Flows exposed near BLM (and inferred for Coyote Ridge) are the farthest outflow along a paleochannel sourced to the ESE. Flows exposed on the Whale (ncWa1 and swWa2-5) appear to have come from the NNW and NNW-WNW, respectively. Flows in the northeast Whale (neWa4) appear to have come from the W, but they are northeast of a fault that separates this subdomain from the rest of the Whale, and the amount of separation is not known. Flows exposed on the Squid appear to have come from the NNE-NE, but back-projection might be challenging. The Whale and Squid were both in the southern part of the field, and the Squid was at the margin of the field. It is possible that flows that went through or near the southern Whale and continued but were redirected to the SSE-SSW by paleotopography. Alternatively, the Whale and Squid might represent different parts of the southern basalt field that have been juxtaposed by the Coyote Lake fault. These projections suggest an eruptive center somewhere W-NNW of the northern Whale, but neither vents nor dikes have been identified, and it is not known if vents were eroded or are buried in an adjacent basin. Other than CRTH1 and CRTH2, there are no other boreholes in the area. The analysis of Miller and Buesch (2022) indicates that dismembered Pliocene and late Miocene basalt fields in the Mojave Desert retain vent structures in several cases, which leads us to conclude that the vent areas were buried.

Geochemistry from across the Bicycle Lake basalt field indicates most lava flows are basaltic andesites and are locally interstratified with basalt. In the south Whale and Squid, most basaltic andesite is A/T TR, and is locally interstratified with A/T HR basaltic andesite and possibly a (slightly altered) basalt. In the north Whale, basaltic andesite is A/T TR (but slightly higher TR than in the south Whale) and is interstratified with basalt with A/T LR. The compositional sequences at north and south Whale began and ended with TR basaltic andesites, but the “excursions” in composition were during the middle of the eruption sequences and differed in the two locations. In Bicycle Lake mesa, the basaltic andesite is A/T TR and HR, and appear similar to lavas from north and southwest Whale, respectively, and would imply two flow paths merged into the single path that delivered lava to the Bicycle Lake mesa area. The distribution of compositions suggests minor (but not complete) sector partitioning of flow paths during the growth of the field. The variations in compositions are not large but are similar to the range in the 5.46 Ma Broadwell Mesa basalt field where the A/T range is 4.4–6.1 (Buesch, 2017). In the Broadwell Mesa field, the A/T increased upward across the field, but some sections have more limited ranges based on incomplete sections (typically the bottom of the section was not exposed) or what lava flowed to the specific areas.

The similarity of the basaltic andesite geochemistry (A/T TR, MgO, Cr, Zr) at the bottom of the CRTH2 borehole with all other samples strongly indicates that it represents the Bicycle Lake basalt, although there are ~10 Ma basalt exposures (with no geochemistry data) >8 km northeast of the Whale (Buesch and others, 2018). Although there are no geochemical analyses for these Miocene basalts, comparison of the Bicycle Lake basalt to Miocene basalts and basaltic andesites in western Fort Irwin shows nearly distinct fields with only one sample in common (Fig. 5). The geometry of the basalt in CRTH2 at a depth 293 m compared to the exposed basalt ~1 km to the south at the north end of Whale indicates significant down to the north and probable sinistral separation along the Bicycle Lake fault. This outcrop-borehole pair will be valuable as a long-term constraint when compared to youthful offset values and kinematic indicators along the fault.

Each of the three largest exposures of the basalt have some form of deformation resulting from fault activity since the field formed. At the Whale, transpressional strain transferred from the Coyote Lake to Bicycle Lake faults is manifest in a series of reverse faults and faulted kink folds along the west side and across the southern domains, and probably results in the broad arching and northward plunging of the basalt in the central and especially northern domains. The curving of fold axes toward the south might also owe to drag along the Coyote Lake fault to the south. The basalt in the Squid and Bicycle Lake mesa areas appear to be arched, and possibly represent transpressional strain where faults change orientation into a constraining bend, or near-fault deformation. Conversely, gaps in these three large basalt exposures represent downfaulted areas or tectonic sags, as demonstrated by boreholes CRTH1, CRTH2, and lack of basalt in Bicycle Lake basin. We conclude that topography in southern Fort Irwin records Pliocene and Quaternary tectonic deformation, as suggested by Miller and Yount (2002).

Thus far, neither unique piercing points nor unique lava flows have been identified for use to constrain separation on faults. The detailed geochemical stratigraphy in the Whale SW subdomains 2, 3, and 4 might be able to constrain separation on the faults, but more samples need to be analyzed. There is still an open question as to whether the Whale and Squid formed in close proximity (minimal separation on faults) or have been juxtaposed along the Coyote Lake fault. This question is important for reconstructing the field because the east-west extent of the Squid is greater than that for much of
the Whale, and the architecture of the flows in the Squid may constrain the geometry of the southern part of the volcanic field. Matching it to the remainder of the field by independent fault separation estimates will provide better constraints for the geometry of the Whale part of the field. Langenheim and Jachens (2014) noted that no offset along the Coyote Lake fault was evident in gravity and aeromagnetic data sets. In particular, a magnetic body, elongate ~N-S, lies east of the Whale and the Squid and is not measurably disrupted at the projected Coyote Lake fault. This observation indicates an estimated <2 km offset on the fault. Accordingly, the folds observed in the Whale may be a manifestation of shortening at the termination of the fault.

Using the premise there was a topographically controlled basal flow to the WNW from an area near the Whale, fault reconstruction can provide estimates of the distance traveled by the flow. Restoring 4.1–4.8 km of left separation on the Coyote Canyon fault (Schermer and others, 1996) places the basalt on Coyote Ridge near the projection of the basal-filled paleochannel at Bicycle Lake mesa. This results in ~6 km between the faults where the basal-filled paleochannel is buried beneath Bicycle Lake basin fill, although along this projection one borehole in southern Bicycle Lake that reached bedrock did not intersect basalt. Restoring 3–8 km of separation of bedrock exposures (Schermer and others, 1996) along the Bicycle Lake fault reduces the 12–16 km gap between BLM and the Whale to 4–13 km. The total fault-restored flow distance from the northern Whale to Coyote Canyon exposures, including the 1.5 km long exposure at Bicycle Lake mesa, is 10–20 km. These flow distances are within the ranges of several of the Neogene basalt fields in the Mojave Desert, whose maximum distance of flow was ~20 km (Miller and Buesch, this volume). It is also notable that several other basalt fields (Broadwell Mesa, Pisgah, and Pipkin) have long breakout flows from the main field. (Buesch, 2017; Miller and Buesch, 2022; Buesch et al., 2022).

Comparing stratigraphic sections of basalt (1–20 m thick) in the different parts of the volcanic field and the distances traveled by the flows indicates high to very high aspect ratios (length/thickness) of the basalt. At the Squid, aspect ratios are 290–4,300 and these are limited by the size of the Squid, and the Whale has aspect ratios of 375–7,500. After removing the possible separation along the Bicycle Lake fault, the aspect ratios from the Whale to BLM are 650–4,000, and the Whale to Coyote Ridge are 4,000–10,000. These distances and aspect ratios indicate that if the paleotopography allowed long runout of basalt flows, then basalts might be buried in basins in or adjacent to the volcanic field. In the Superior basin of western Fort Irwin, basalt flows in boreholes are from the Black Mountain volcanic field to the west, and the distance to the edge of the exposed flows of the field is ~8 km, but because the eruptive center for the field has not been identified, the distance of flow might have been ~25 km (Buesch, 2018). Based on basal in boreholes in Superior basin and in the Harper Lake basin to the west of Black Mountain, Miller and Buesch (2022) infer a minimum areal extent of 630 km² for Black Mountain field. Not only can these long runout basal flows be stratigraphic markers in basins, Buesch (2018) concluded that because basalt flows (1) are crystallized rock with cooling joint fracture systems, (2) have long runout, and (3) are continuous from surface exposures (which are possible water recharge sites) to deposits in sedimentary basins, they might exert an important control on the recharge and storage properties of groundwater in the basins.

Pliocene paleogeography has been disrupted by the ECSZ faults; however, the distribution of the basalt can provide insights into regional drainage. The Coyote Ridge area is highly faulted, and basalt was deposited on upper Miocene to Pliocene(?) sedimentary rocks (Schmer and others, 1996, Figure 4); however, there are also Jurassic(?) plutonic and metamorphic rocks in the area, so the depth of the sedimentary basin is not known. The basalt at Bicycle Lake mesa was deposited on Jurassic(? plutonic and metamorphic rocks (Miller and others, 2014), a possible bedrock sill between basins. To the southeast near the Whale and Squid, the basalt was deposited on a thick sedimentary sequence. Along the eastern side of the Whale there is >250 m of pre-basalt sedimentary rock exposed, and in CRTH1 there is 115–320 m of pre-basalt sedimentary rock. South of the Squid, the basalt was deposited on >220 m of upper Miocene sedimentary rocks, and the sequence may total ~1 km. Another disruption to the Pliocene paleogeography is the disruption and reversal of the paleochannel along which the basalt flowed. In the original paleochannel, the basalt flowed down the topographic slope from the vicinity of the Whale, across the area at Bicycle Lake mesa, and to the Coyote Ridge location. With an assumed 1° slope on the paleochannel, there would have been ~430 m of relief from the Whale to the Coyote Ridge area. The modern elevation of the basalt at Coyote Ridge is at ~1,050 m, the highest elevation for any of the basalts, and is a maximum of ~650 m at the Whale. This elevation difference of 400 m plus the relief needed for flow to Coyote Ridge yields a change of ~830 m, a reversal of relief over ~25 km distance during the last 4.5 m.y.

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Southwest Nevada and Lunar Crater volcanic fields: similarities and differences and ideas for more research in the desert

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ABSTRACT—Southwest Nevada and Lunar Crater volcanic fields are two well-studied basaltic systems in Nevada, separated only by about 120 km. While the lavas that they erupted during the past few million years had similar compositions, they differ in their mantle sources, total volumes and numbers of volcanoes, and in some aspects of their eruptive styles. They also provide excellent opportunities to study geomorphic evolution of volcanic surfaces in arid settings.

Southwest Nevada volcanic field

Southwest Nevada volcanic field (SNVF; Figure 1) is located about 120 km northwest of Las Vegas, about 40 km from Death Valley. The field began its early activity about 15 million years ago in the Miocene epoch with several large caldera-forming eruptions, but after about 10 million years ago it has been dominated by small-volume mafic volcanoes. During the past five million years about 17 small volcanoes erupted (Valentine and Perry, 2007), with compositions of basalt, trachybasalt, and basaltic trachyandesite. All of these are monogenetic, meaning that each only had a single eruptive episode and was then permanently dead; note that an eruptive episode may last for weeks to decades and can have quite a complicated range of eruption processes during that time. The volcanoes have progressively decreased in size, with a single edifice at 2.5 km$^3$ at about 4.6 Ma to ~0.1 km$^3$ and smaller for the Quaternary volcanoes. The youngest, Lathrop Wells volcano, is about 77,000 years old. This relatively young age and its proximity to the proposed Yucca Mountain radioactive waste disposal site is why Lathrop Wells and the other Quaternary volcanoes in the area received much attention during the 1990s and early 2000s.

There are eight Quaternary volcanoes (ages <2.588 Ma) in SNVF, forming a north-northwest trending band about 60 km long and 10 km wide. Lava flows typically extend between 1–1.5 km from their source cones. The lavas are dominated by 'a', which can be related to relatively high effusion rate, viscosity increases due to degassing during shallow magma ascent, and other factors. Nearly all of the lava fields contain “rafts” of scoria—these are chunks of the source cones that failed and were carried away atop the lavas. The scoria cones range from ~30-100 m tall, and are increasingly degraded by weathering and erosion with age (Valentine et al., 2006, 2007). Lathrop Wells volcano is sufficiently young that its fallout tephra deposit is preserved. This deposit extends up to 20 km from the cone and suggests that Lathrop Wells exhibited an eruption style known as violent Strombolian, which involves sustained eruptive plumes up to ~10 km tall that spread downwind and drop scoria lapilli and ash over the landscape. Such eruptions are driven by magmatic volatiles, mainly H$_2$O. Unlike many basaltic magmas, which can have H$_2$O contents of <1 wt%, Lathrop Wells likely had ~5 wt.% water vapor, which might have aided in the development of these eruptive plumes (Nicholis and Rutherford, 2004; Valentine et al., 2007).

During the Quaternary, SNVF has produced eruptive material (lava and scoria) at a long-term rate of about 0.5 km$^3$/Myr, a truly miniscule rate by the standards of most volcanic systems. The time between formation of new volcanoes seems to depend upon the size of the previous eruption; for example, formation of a relatively large cone and lava field (by SNVF standards) is followed by relatively long repose time before the next scoria cone forms. This time-predictable behavior may be related to stress relief in the crust. As the area is slowly extending, larger dikes,

Figure 1. Relief map of the southwestern United States showing locations of the Southwest Nevada and Lunar Crater volcanic fields, and the region of the Mojave Desert for reference.
which feed large cones, will relieve more stress than smaller ones. Isotopic (Figure 2) and other geochemical data are consistent with the magmas having been sourced in the solid uppermost part of Earth’s mantle (lithospheric mantle), and are likely related to small pockets of volatile-enriched mantle that are slightly molten under ambient conditions at about 70–90 km depth. Valentine and Perry (2007) hypothesized that the very small portions of partial melt are passively mobilized by tectonic strain, forming dikes that ascend quickly to the surface with little or no residence time in the crust.

**Lunar Crater volcanic field**

Lunar Crater volcanic field, to the northeast of SNVF (Figure 1), is also in an area that experienced early large-scale caldera-related volcanism (Miocene), but during the past ~5 million years volcanism has been nearly entirely of basaltic, trachybasaltic, and tephritic composition, generally similar to SNVF. In contrast to SNVF, the Lunar Crater field has produced at least 161 monogenetic volcanoes during this time (Valentine et al., 2017), with volumes typically between 0.1–1 km$^3$. Volumes of individual volcanoes do not show any clear trend with time. The youngest activity was ~35,000 years ago during formation of the Marcath volcano with its cone, rampart, and lava field (Shepard et al., 1995; Valentine et al., 2017).

There are at least 96 Quaternary vents in the LCVF, forming a north-northeast trending band about 30 km long and 5 km wide (Tadini et al., 2014; Valentine et al., 2017). Clearly the spatial density of Quaternary volcanoes in LCVF is much higher than the SNVF with its eight volcanoes spread over a length of 60 km. Lavas in LCVF are also dominated by ‘a’a morphology, and rafted cone pieces are quite common, recording an interplay between cone building and destruction by lava flows (Younger et al., 2019). Cones range in height from ~50 to 200 m, and lavas extend up to ~5 km from the source cones; this likely a result of both greater volumes of magmas and higher supply rates during individual eruptions, compared to SNVF. Although there are many tephra fallout beds exposed in quarries and natural outcrops, only that which is associated with Marcath volcano can be clearly traced and studied; it extends about 5 km downwind (east-northeast) from the cone, and records an eruption plume up to ~7 km high with an eruptive style akin to typical Hawaiian eruptions such as the 1959 activity at Kīlauea Iki (Johnson et al., 2014). Water contents of the magmas were indirectly estimated to be in the range of 5–7 wt%, similar to SNVF (Cortés et al., 2015). Unlike SNVF, the Lunar Crater field contains about four maar volcanoes, which record violent explosions caused by interaction of magma and groundwater; one of these is Lunar Crater itself (Valentine et al., 2011). Three of these maars formed during complex eruptive episodes that also featured more “normal” Strombolian and Hawaiian styles of eruptions (Valentine and Cortés, 2013; Amin and Valentine, 2017).

It is difficult to measure individual eruptive volumes of the Quaternary LCVF volcanoes because many of the lavas, tephra, and even cones are partly to completely buried by products of subsequent eruptions and/or are buried beneath alluvial sediments in subsiding basins. A very rough estimate is that the Quaternary activity has resulted in ~10 km$^3$ of eruptive products (this has a huge uncertainty range), which would equate to ~4 km$^3$/Myr for the long-term magma output rate. This is roughly an order of magnitude larger than that of the Southwest Nevada field. The age and volume uncertainties are too large at LCVF to test whether the system there is time predictable. Unlike SNVF, the Lunar Crater magmas were sourced in asthenospheric mantle (e.g., Figure 2), i.e., the convecting part of the upper mantle (just beneath the lithospheric mantle), likely due to upwelling during regional extension. Geochemistry suggests that this asthenospheric mantle is heterogeneous in the details of its composition at scales as small as several hundred meters, based upon differences between adjacent cones. In addition, upwelling brought different mantle domains into the melting region as time progressed; these domains record the long memory of the mantle such as contributions from ancient subducted ocean crust and the fluids it introduced (Rasoazanamparany et al., 2015; Valentine et al., 2017). These fluids, along with a small degree of partial melting of source mantle, contributed to the relatively high (for basaltic magmas) estimated water content of the magmas. The gradual changes in mantle sources, which are reflected mainly in isotopic composition through time, allowed us to identify four
magma genesis episodes during the ~5 Ma of volcanic history at LCVF (Valentine et al., 2017).

Opportunities

From a broader perspective of research in desert environments, volcanic fields such as Southwest Nevada and Lunar Crater have a wide range of volcano ages and formed by a range of eruptive processes (Figure 3). Thus, the fields provide excellent opportunities to constrain rates of geomorphic processes and how those might depend upon the original characteristics of a surface (clast size, texture, degree of induration, slope). There was a substantial effort made to use basaltic volcanoes as geomorphic laboratories in the 1980s, focused at Cima volcanic field in the Mojave Desert (e.g., Wells et al., 1985; McFadden et al., 1986; Dohrenwend et al., 1986, 1987). While many key results came from this work, it was hampered by reliance on the K-Ar dating method (Dohrenwend et al., 1984), which is often not accurate for young basaltic rocks. Since those important studies, the $^{40}$Ar/$^{39}$Ar method and cosmogenic surface exposure age dating methods have come to the fore and advanced in their accuracy, which would allow better age control on volcanic surfaces. Additionally, a limited amount of research indicates that the initial characteristics of volcanic surfaces, which can be reasonably evaluated even for older Quaternary volcanoes, can play a key role in the rate at which a surface evolves (Figure 4; Valentine and Harrington, 2006; Valentine et al., 2006, 2007).

Figure 3. Two volcanoes of different age in the Lunar Crater volcanic field. (A) View looking north toward the ~35 ka Marcath cone and its lava field, which extends westward from the Pancake Range into Big Sand Springs Valley. The cone is pristine, with only small degrees of gully development and creep features such as garlands, and the lava also preserves its original surface morphology, texture, and topographic-low-filling character (Valentine et al., 2017; Younger et al., 2019). (B) View looking north toward the Pliocene Dark Peak volcano in the Reveille Range (volcano is not dated, but likely is between 4–6 Ma; Harp and Valentine, 2015). The cone is nearly completely eroded away, with only welded near-vent pyroclastic deposits and internal dikes preserved at Dark Peak. The lava field extends westward into what was a valley at the time of eruption, but now forms inverted topography as the lava formed a caprock while the surrounding terrain has eroded. In a field such as LCVF all stages of geomorphic evolution of cones, lava surfaces, and tephra layers may be exhibited. Figure modified from Valentine et al. (2017).

Figure 4. Google Earth image of Red Cone volcano (~1 Ma) in Crater Flat, part of the Southwest Nevada volcanic field. Only a remnant of the scoria cone is preserved. Even though the two lava fields are the same age, their surfaces have evolved differently. W lava field has abundant mounds of rafted scoria cone material, providing both surface relief and sources of loose clastic material, and has advanced development of gullies on its surface. E lava field has only ~3 rafts on it, which have developed gullies, but the remainder of the flow field has essentially no gullies. The differences between the two lava field morphologies led some researchers to conclude that the lavas had substantially different ages, but their differences are due to the initial starting characteristics (Valentine et al., 2006).
Much remains to be learned about the volcanology, geochronology, and petrogenesis of basaltic monogenetic volcanoes in arid to semi-arid environments of the southwest United States (Valentine et al., 2021). The combination of quantitative volcanology and geomorphology, along with application of geochronological techniques that have greatly improved our ability to date young basalts, is a particularly interesting area of research to pursue, and which might lead to new insights into desert geomorphology more broadly.

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Age(s) and characteristics of Saddleback Basalt: Implications for the Miocene stratigraphy and origin of the Boron deposit, Mojave Desert, California

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Abstract—A field, petrographic, and geochronologic investigation of mafic volcanic rocks previously mapped as the Saddleback Basalt in the vicinity of the Rio Tinto borax deposit at Boron, California, sheds important new light on the local eruptive history, age of mineralization, and Miocene stratigraphic framework. Exposures in the Kramer Pit, at Saddleback Mountain and Stonehouse and Murok Hills, and in drill holes east of the Kramer Pit (east Boron basin) demonstrate that these mafic rocks include both lava flows and intrusive sills and that they occur at several levels within Miocene stratigraphic sections. Petrographic characteristics and new $^{40}\text{Ar}/^{39}\text{Ar}$ geochronologic data confirm this polyphase eruptive history and indicate that mafic rocks previously lumped as Saddleback Basalt (SB) fall into three distinct compositional and age groups:

1. **“True” Saddleback Basalt**: These 20.5 ± 0.3 Ma coarse-grained (diabasic) plag-olvine-cpx basalt lava flows commonly have a well-developed diktytaxitic texture and include the “type” SB at Saddleback Mountain. Multiple flows occur near the base of the Miocene section, resting directly on Mesozoic basement rocks or on rhyolitic pyroclastic and epiclastic deposits.

2. **Stonehouse Andesite**: These 19.4 to 19.7 (± 0.2) Ma fine-grained, nearly aphyric andesite lavas with 1–2% small (<200µ) olivine phenocrysts in a plag-rich trachytic groundmass are widely exposed in the Murok Hills to the west and north of the Boron deposit, and are intersected by many of the drill holes in the east Boron basin. They include at least 4 to 5 flows intercalated in the lower parts of the lacustrine sections in the Boron basin, or resting directly on older SB.

3. **Boron Basaltic Andesite**: These ~19.2 ± 0.2 Ma basaltic andesite lavas and compositionally identical sills contain obvious phenocrysts of plagioclase and olivine in a fine-grained plag-rich groundmass. Multiple lava flows separated by vesicular rubble horizons are exposed on the north wall of the Kramer Pit and in at least one drill hole. Sills of this age are common in the lacustrine shale facies throughout the Boron basin, and one of these sills is extensively exposed at the bottom of the Kramer Pit, where it directly underlies the main borate ore horizon. The sills are distinguished from flows by their quenched, knife-sharp irregular upper and lower contacts. Extensive soft sediment deformation adjacent to sill margins and proximity to overlying coeval lava flows suggests these sills were emplaced into wet sediments only a few meters beneath the lake bottom.

Distinct ages and mineralogical characteristics of the three mafic rock groups permit better correlation among drill holes, surface exposures, and mine workings. True Saddleback Basalt flows are only exposed close to the Saddleback Mountain area, whereas the Murok Andesite and Boron Basaltic Andesite flows and sills are widespread in the Boron basin and hills to the north and west. The Miocene section of lacustrine deposits and overlying arkoses in the Boron basin generally thickens southward towards the E-W striking, north-dipping basin-bounding West Boron fault. At least one significant intrabasin NW-striking fault separates the east Boron basin into a northern section with abundant 19.2 to 19.5 Ma flows and sills, and a southern section that lacks them. This fault likely has a dip-slip (NE side down) component and several kilometers of dextral strike-slip displacement.

Borate mineralization appears to be closely associated (spatially and temporally) with the emplacement of the 19.2 Ma basaltic andesite sills, as it occurs mainly in claystones directly above the sills, and the age of mineralization is tightly bracketed by ash layers to have occurred ~ 19.2 Ma. The combined stratigraphic and structural relations support a general model for borate mineralization that includes the following elements: (a) Development of an asymmetric (half-graben) Miocene extensional basin, beginning ~19.5 Ma, bound on the south by a N-dipping normal fault, with lacustrine deposits...
accumulating syntectonically in the proximal part of the hanging wall; (b) Eruption and sub-lacustrine emplacement of extensive basaltic andesite sills into wet lake sediments just below the lake floor at ~ 19.2 Ma, accompanied by the development of a sub-lacustrine hydrothermal system/hot springs and possible changes in the chemistry and/or temperature of the lake waters, promoting precipitation of borates; (c) Subsequent burial of these borates beneath additional lacustrine shale and progradation of arkosic fanglomerate and sandstone covered and preserved the borates. The ultimate source of the boron-rich magmatically driven hydrothermal fluids is speculative, but might have involved Miocene hydrothermal leaching of boron from underplated marine sediments of the Rand Schist inferred to underlie the region in the shallow subsurface.

**Introduction**

The world class borate deposit near Boron, CA in the Mojave Desert supplies most of the United States’ boron—used in everything from fertilizers to borosilicate glass. Despite its economic importance, the regional tectonic setting of this deposit, the source of the boron, the nature of the hydrothermal system that introduced the borates, and the relationship (if any) of borate mineralization to local magmatic activity remain poorly understood.

Scattered exposures of mafic volcanic rocks in the vicinity of the deposit and within the mine workings (Kramer Pit) (Fig. 1) were collectively named “Saddleback Basalt” by Gale (1946) and assumed to be a single unit, a conclusion followed by subsequent workers (e.g. Dibblee, 1967; Armstrong and Higgins, 1973; Golombek and Brown, 1988; Siefke, 1991). Dibblee (1967) formally named the Saddleback Basalt (SB) as the middle member of the Tropico Group, a sequence of Miocene volcaniclastic, lacustrine and alluvial/fluvial sedimentary rocks that are widespread in the western Mojave Desert region.

Because borate mineralization in the Kramer deposit consistently lies stratigraphically above SB and is enclosed within the overlying lacustrine shales of the upper member of the Tropico Fm, it was generally assumed that SB basaltic volcanism predated and was largely unrelated to borate mineralization. Indeed, exploration efforts have focused mainly on sections above the SB, and drilling is generally stopped once the first mafic volcanic unit is reached.

This study examines the many different occurrences of what had previously been designated as Saddleback Basalt in the vicinity of the Boron mine, including exposures in the Kramer Pit, Saddleback Mountain, Stonehouse and Murok Hills, unnamed hills to the N and W of the mine, and in drill core from a large number of holes drilled in the East Boron Basin area (Fig. 1). Samples collected from most of these sites were analyzed petrographically, and a large subset were then radiometrically dated. The analytical work confirms what was clear from field relations—these mafic volcanic rocks are not all the same age or composition, but instead record several distinct pulses of mafic volcanic activity in the area.

Furthermore, some of the mafic rocks previously assumed to be extrusive lavas are actually shallow sills emplaced into poorly lithified lacustrine deposits. A new stratigraphic nomenclature is proposed for volcanic rocks previously lumped as the Saddleback Basalt and I explore the implications of these new field and analytical observations on the structure, stratigraphy, and origin of borate mineralization in the Boron basin. Intriguingly, it appears that the youngest of these pulses of mafic magmatic activity appears to...
be closely associated with borate mineralization, with implications for future exploration efforts.

**Regional geologic setting**

The Kramer Deposit, site of the Rio Tinto US Borax open pit mining operation, is located north of the town of Boron in the Mojave Desert. The area is characterized by broad pediment surfaces covered by a veneer of alluvial gravels, with isolated hills exposing diverse bedrock units including Mesozoic granitoids and Miocene volcanic and sedimentary successions. In this part of the Mojave, the Miocene successions have been generally assigned to the Tropico Group of Dibblee (1967), originally considered to be Pliocene, but re-assigned to the Miocene following identification of fossil mammal remains (Whistler, 1984) and K-Ar dating of intercalated volcanic rocks (Armstrong and Higgins, 1973). Dibblee (1967) divided the Tropico Group into three members: a lower member dominated by reworked rhyolitic tuffs, volcaniclastic sandstones, and lacustrine deposits, a middle member of mafic lavas he named the Saddleback Basalt, and an upper member of lacustrine claystones, sandstone, and arkose. Implicit in this stratigraphic framework was the assumption that the scattered mafic volcanic rock exposures were all correlative and approximately the same age. The Tropico Group is assumed to be broadly correlative with Miocene successions to the west (e.g. Gem Hill Formation) and to the east in the vicinity of Barstow (e.g. Pickhandle and Barstow formations). Less clear is whether these formations represent distinct basins, and if so, how large and how many different basins are represented.

The tectonic setting of these Miocene basin successions is controversial. Did they accumulate in fault-bound extensional basins (Dokka et al, 1991), flexural sags (e.g. Fillmore et al, 1994), or perhaps as infilling within topographic lows between older structural highs or Miocene volcanic edifices? It is likely that there was significant early Miocene extension in this region, but the precise timing and magnitude are uncertain, as is the related issue of how much of the exposed Tropico Group is pre-, syn- or post extensional. The precise geometry and controls on Miocene sedimentation patterns are difficult to assess due to limited exposure and extensive Quaternary cover. The modern topography pattern is of little help, as it is largely a consequence of either differential erosion or, more commonly, localized late Miocene to Recent uplift and transpression along segments of the NW-striking dextral strike-slip faults that cut across this portion of the Mojave Desert—part of the eastern California shear zone (Dokka and Travis, 1990). A more complete understanding of the Miocene stratigraphic framework of the central and western Mojave region is going to require much better age control and correlation of exposed sections, improved subsurface (geophysics and drilling) information on what underlies the extensive alluvial cover, and more detailed stratigraphic information (facies analysis, paleocurrents, etc) of the rock units. This study helps advance those goals somewhat by providing important new insights on the origin, geometry and evolution of one of these Miocene basins—the Boron basin.

**Geologic framework of the Boron area**

The general geologic framework of the Boron area was first established by the pioneering mapping efforts of Dibblee (1958a, 1958b) with additional information on

![Figure 2: Highly simplified geologic map of the Boron mine area showing present outline of Kramer Pit, drill hole locations in East Boron basin, major structures and rock units, and locations of selected radiometric ages. Compiled from Dibblee (1967), US Borax, (unpubl. mapping, 2005), and field work and image analysis by the author.](image)
the Miocene stratigraphy provided by Dibblee (1967). Bedrock is exposed (from west to east) in the Murok Hills, Stonehouse Hills, in the Kramer open pit (including unnamed hills surrounding the mine), and in the Saddleback Mountain area to the north and northeast of the mine (Fig. 1) Additional information is provided by drill core from a dense array of deep holes that were drilled by US Borax in the east Boron basin area (Fig. 2) and is summarized below. Basement consists primarily of deeply weathered Mesozoic granitoids of various types. Overlying the basement unconformably are up to 1 km-thick successions of Miocene sedimentary and volcanic rocks of the Tropico Group. The sedimentary rocks of the Tropico Group in this area include a lower interval of rhyolitic tuff breccia, tuffaceous sandstones and siltstones, and minor lacustrine limestone, a middle interval dominated by lacustrine claystones and siltstones, and an upper interval dominated by arkosic sandstone and fanglomerate, but there are considerable lateral variations in both lithology and thickness. This is not surprising given the inferred depositional environments of isolated desert lakes surrounded by subaerial fans and fluvial systems. An intercalated primary pyroclastic flow near the base of the Miocene section at Saddleback Mountain yielded a $^{40}{\text{Ar}}/^{39}{\text{Ar}}$ age of $21.0 \pm 0.4 \text{ Ma}$, whereas reworked tuff horizons within the lacustrine claystone middle sequence that includes the borate horizons yielded ages of $19.3$ to $18.8 \text{ Ma}$ (Kozak, 2000, see below). The upper arkoses are undated. As described below, the intercalated mafic volcanic rocks in the Boron area that were previously grouped as Saddleback Basalt are of diverse compositions and occur at different stratigraphic levels, yielding ages that range from ~18.5 to 21 Ma.

The geology of Kramer deposit has been investigated in detail by several authors (Gale, 1946; Barnard and Kistler, 1965, Diblee, 1967; Siefke, 1991) and is summarized below. The lenticular stratiform borate horizon is enclosed within lacustrine claystones that are part of what locally is referred to as the Kramer Beds (Fig. 3). This conformable section, broadly correlative to Dibblee’s (1967) upper Tropico member and superbly exposed in the mine working, consists in ascending order of: (a) “Saddleback Basalt” – actually a mafic sill complex erroneously correlated with SB, (b) a thin barren lower claystone interval, (c) the borate ore horizon, zoned from an outer ulexite and colemanite facies to an inner sodium borate facies of borax and kernite, (d) an upper barren claystone interval, and (e) a thick section of arkosic sandstone (Fig. 3). The borate is interpreted to have been syngenetically precipitated directly on the lake bottom, presumably fed by thermal springs highly enriched in boron and sodium. $^{40}{\text{Ar}}/^{39}{\text{Ar}}$ dating of a tuff within the main ore horizon suggest that borate mineralization occurred at $19.28 \pm 0.08 \text{ Ma}$ (Kozak, 2000).

The Kramer beds form a southward thickening wedge that terminates southward at the West Boron fault, a steeply north-dipping normal fault. This east-striking fault separates the mineralized lacustrine section in the mine from granitic basement to the south (Fig. 2) and is hypothesized to have been the basin bounding fault that was active during deposition of at least some of the Kramer beds. Detailed mapping within the mine combined with abrupt changes between sections in adjacent drill holes reveals additional deformation of the Kramer beds in the form of NW-striking faults and related folds, some of which may have been active during lacustrine deposition and mineralization (Barnard and Kistler, 1965).

Methods

Field work included several visits to the Kramer Pit in consultation with US Borax geologists, logging of drill core from the east Boron basin area, and the examination and sampling of surface outcrops in surrounding bedrock exposures at Saddleback Mountain, Stonehouse and Murok Hills, and unnamed

Figure 3 Generalized stratigraphic column for the upper Tropico Group (Kramer Beds) in the Kramer Pit, showing stratigraphic positions of dated samples. Modified from Barnard and Kistler (1965), Siefke (1991), and Kozak (2000)
hills to the north and west. Approximately 40 new samples were collected of potentially dateable units, focusing on mafic volcanic rocks correlated with the Saddleback Basalt. This work builds on a previous reconnaissance geochronological effort undertaken by Gans and his MS student Ivan Kozak (Kozak, 2000) that was aimed specifically at dating borate mineralization. Samples for dating were collected from several reworked tuff horizons within the Kramer Pit, a pyroclastic flow deposit near the base of the section at Saddleback Mountain and from mafic lava flows and sills throughout the study area. Mafic lavas were sampled from the interior portions of individual lava flows, as experience has shown that the best data is obtained from the more slowly cooled, holocrystalline interiors, rather than the quenched (glassy) margins. A detailed petrographic study of these samples was used to characterize the mafic rocks and determine which were most suitable for dating.

Pure feldspar and biotite separates were obtained from the rhyolitic tuffs using standard crushing, sieving, and magnetic and density mineral separation procedures. Mafic lava samples deemed suitable for dating were lightly crushed and sieved to varying size fractions (100–200µ to 300-500µ) and ultrasonically cleaned in de-ionized water. Standard magnetic separation techniques and hand picking were used to generate groundmass concentrates. Splits of each sample (ranging from 1 to 30 mg) were then encapsulated in copper packets and loaded into a sealed quartz vial interspersed with packaged flux monitors. The samples were irradiated in a cadmium-lined tube at the TRIGA reactor at Oregon State University for 14 hours. Samples were then analyzed by incremental heating in a Staudacher-type resistance furnace using the general procedures and system described by Gans (1997). Analyses included total fusions on single grains or small aliquots and step heating experiments for each sample. The mafic volcanic (groundmass) samples commonly yield complex age spectra due to the combined effects of reactor induced recoil, low temperature argon loss during weathering, and excess argon, as discussed below. All errors given for our estimated (preferred) ages in this report are ± 2 sigma (estimated 95% confidence). The flux monitor used for all irradiations was Taylor Creek Rhyolite with an assigned age of 27.92 Ma (Dalrymple and Duffield, 1988). For comparison, we obtain an age of 27.75 Ma on Fish Canyon Tuff sanidine (another widely used standard) when we use 27.92 Ma Taylor Creek Rhyolite as our flux monitor.

Field observations on “Saddleback Basalt” in the Boron area

“Saddleback Basalt” in the Kramer Pit

Rocks mapped as Saddleback Basalt occur along the northern and northwestern wall of the Kramer Pit as well as in the bottom of the pit. On the upper benches of the northern wall, we examined an interval of olivine-plag phryic mafic lavas that dip gently southward. At least 4 or 5 individual lava flows were identified based on the presence of vesicular tops and rubbly oxidized bases. Samples 04PGMJ-67 and 04PGMJ-68 were collected from the uppermost and 4th flow down respectively. The basalt flows are overlain directly by clay shale, but the position of the lavas relative to the Saddleback Basalt and ore horizon to the south is not entirely clear. The most likely scenario is that these represent lavas erupted from a vent to the north that flowed southward, stalling at the ancestral lake shoreline.

A massive body of basaltic andesite is exposed in the bottom of the Kramer Pit, a few m below the main borate horizon. This interval is at least 30–40 m thick, and grades rapidly downward from a quenched, fine-grained upper contact with the overlying claystone to a slightly coarser grained massive interior with visible plagioclase, olivine, and clinopyroxene phenocrysts. A number of features along the upper contact indicate that this body is a sill rather than a lava flow. The upper part of the basaltic andesite is quenched, and although vesicles are present, the upper contact lacks the highly vesicular layering and rubble that is characteristic of most mafic lava flows. Instead, the contact is knife sharp, somewhat irregular and locally cuts beds, and the overlying claystone within a few feet of the contact appears baked. In addition, abundant soft-sediment deformation and disharmonic folding of the claystone immediately adjacent to the contact suggests that bedding was disrupted by intrusion of the body. The interior of the basaltic andesite is massive, with no obvious flow fabric, and a grainy texture typical of shallow mafic intrusives. Similar relations in many of the drill holes to the east (described below), suggest that this is one of the larger of many sills.

“Saddleback Basalt” at Saddleback Mountain

The type section of the Saddleback Basalt is at Saddleback Mountain, ~ 5 km ENE of the Kramer Pit. Here, basalt occurs in two bands: one on top of Saddleback Mountain, and a parallel strip to the southwest (Fig. 2). In both areas, the basalt consists of coarse-grained (diabasic) olv-plag-cpx phryic basalt with abundant vesicles and a conspicuous diktytactic texture. Several lava flows are present as indicated by oxidized highly vesicular rubble horizons separating the more massive interiors of adjacent flows. Both sections dip moderately southwest and are separated by a thin strip of Mesozoic basement overlain by reworked tuffaceous sediments and a thin welded ignimbrite dated at 21.0 ± 0.4 Ma by Kozak (2000). The southern strip of basalt clearly overlaps this tuff with modest angular unconformity. The simplest explanation of the map pattern is that the two parallel bands of basaltic flows have depositional contacts on basement (or older pyroclastic deposits) on their northeast margin, but have been repeated by a NE-dipping normal fault that cuts across the southern flank of Saddleback Mountain.
Additional exposures of “Saddleback Basalt” were examined in the dark hills approximately 2 km to the NW of Saddleback Mt. Here, a basal lava of diabasic basalt indistinguishable from the one at Saddleback Mountain is directly overlain by an assemblage of lavas and breccias of a much finer grained, brown to reddish weathering nearly aphyric andesite. The preserved remnants of a cinder cone of the same andesitic material suggests that at least some of it was sourced from this area.

“Saddleback Basalt” in the Stonehouse and Murok Hills
Exposure is very poor in most of these hills, with resistant bands and caps of mafic volcanic rubble overlying recessive slopes underlain by float of light-colored sedimentary rocks (Fig. 1). The best outcrops of mafic rocks appear to be conformable and intercalated with intervals of thin bedded tuffaceous sandstones, siltstones, and claystones, and all have a very different character than the basalt at Saddleback Mountain. Most exposures consist of very fine grained to glassy, brown-weathering lavas and sills(?) that lack obvious phenocrysts (i.e. are aphyric) or very crystal poor, with a trace amount (< 2%) of iddingsitized olivine microphenocrysts. A few have phenocrysts of plagioclase and clinopyroxene, but their relative stratigraphic position is unclear. The better outcrops display a conspicuous platy parting that is more characteristic of andesite than basalt. Some of the exposures are clearly lavas separated by vesicular breccias; others are more likely sills, in that they form discontinuous but roughly concordant sheets with knife sharp contacts. Overall, the poor exposure and subdued relief greatly hampers any attempt to work out the local stratigraphy and assess exactly how many different mafic volcanic intervals are present.

“Saddleback Basalt” in drill core
Crucial insights into the character and relative ages of Miocene mafic volcanic and intrusive rocks from the Boron area were obtained by examining drill core from the east Boron basin area. A total of 13 holes were drilled to depths of 1000 to 2000 ft in an area 1–3 km east of the Kramer Pit (Fig. 2). Simplified drill logs (Fig. 4) demonstrate that this area is underlain by thick sections of Pliocene to Miocene arkosic fanglomerate and sandstone which, in turn, overlie varying thicknesses of Miocene lacustrine deposits and volcanic rocks. Several drill holes bottom in Mesozoic basement, and 7 of the 13 drill holes intersected one or more intervals of mafic volcanic

Figure 4: Generalized columnar sections for some drill holes in the East Boron exploration area, showing age determinations for selected samples collected from drill core. Note the consistent stratigraphic order of three mafic rock types, all previously correlated with Saddleback Basalt, and the abundance of mafic sill associated with the youngest group.
rocks that all broadly correlated with Saddleback Basalt. My observations and sampling focused on these mafic volcanic rocks intervals. The complete sections provided by this drill core made it possible to identify multiple units and their relative stratigraphic position and to assess whether the mafic bodies are intrusive (sills) or extrusive (lava flows) based on their contact relations and internal textural variations.

The presence of multiple intervals of mafic volcanic rocks separated by thick intervals of lacustrine deposits in several of the drill holes (EBC-7, 8, 9, 10, and 11) provides the most direct evidence for multiple episodes of mafic volcanism in the Boron area (Fig. 3, 4). The fact that volcanic rocks at different stratigraphic intervals within the same drill hole had notably different textural and modal characteristics strengthens this argument and suggests that they likely came from different sources. The drill core also provided compelling evidence that at least some of these mafic bodies are sills rather than flows. The sills are distinguished from flows in the drill holes by (a) symmetric quenching of upper and lower contacts as indicated by decreasing grain size and crystallinity toward the margins, and the presence of knife-sharp, irregular contacts between the massive igneous rock and the adjacent claystones, (b) lack of vesicular rubble horizons that characterize upper and lower boundaries of most lava flows, (c) apparent baking and oxidation in directly overlying claystones, and (d) disruption of layering and soft-sedimentation deformation of claystones immediately adjacent to the sill margins.

Sills were identified in 7 of the drill holes and range in thickness from a few feet to 130 feet. In some holes, multiple sills were identified based on the presence of internal quenched margins or intervening thin septa of claystone. All of the sills are similar in terms of their textures and visible crystals of olivine and plagioclase (see petrographic/age groups below). Although sills occur both above and below lava flow sequences in various holes, they always differ in their phenocryst assemblage from underlying flows (e.g. EBC -8, 9, and 10), but closely resemble the overlying sequence of flows observed in EBC-09 (Fig. 4). The lava flows were divided into 3 groups based on their stratigraphic position and general textural and modal compositions (see petrographic/age groups described below). The youngest sequence is observed in EBC-9 and contains obvious plag and olivine phenocrysts. A somewhat older group of at least 4 or 5 individual lava flows was identified lower in many of the drill holes (EBC-7, 8, 9, 10, 11, and 13), and consists of very fine-grained, phenocryst-poor (olivine) andesite. An even older coarse-grained basalt was identified at the bottom of EBC-10, where it is overlain by the fine-grained andesite of the intermediate age group.

Thus, observations from the drill core indicate that multiple episodes of mafic volcanism in the east Boron area produced three distinct groups of lava flows, and also resulted in the widespread emplacement of sills into the lakebeds – most likely associated with the youngest group of flows. In EBC-9, the sills are separated from correlative (?) overlying lava flows by only a few meters of claystone, suggesting the sills were emplaced into the wet lake sediments only a few meters below the lake bottom. Samples collected from surface exposures in nearby hills and in the Kramer Pit tell a similar story. Careful petrographic examination of these samples shows that they fall into the same three compositional/petrographic groupings as was seen in the drill core, and wherever the relative stratigraphic position could be assigned, they have the same relative ages.

\[ \text{40Ar/39Ar geochronology and petrographic observations} \]

Overview of \[^{40}\text{Ar}/^{39}\text{Ar}\] dating of mafic volcanic rocks

Dating fine-grained mafic volcanic rocks using the \[^{40}\text{Ar}/^{39}\text{Ar}\] system is a complicated task. Potassium is not a major component of phenocryst phases, so it is concentrated in late-crystallizing groundmass minerals, or in the interstitial glass of the groundmass. Radiogenic argon is commonly lost from glass during weathering, and phenocryst phases can contain significant amounts of non-radiogenic (excess) \(^{40}\text{Ar}\). Additionally, recoil of \(^{39}\text{Ar}\) produced by the irradiation of very fine-grained crystalline aggregates in the groundmass can affect the results.

Most of the samples analyzed in this study yielded “disturbed” age spectra, with old apparent ages for the lowest temperature steps, and ages that decreased monotonically until approximately 50–60% of the gas was released, at which point the spectra tended to flatten out (Fig. 5). Individual spectra range from “hump-shaped” to “U-shaped” to “L-shaped”—a range of spectral types that is typical of fresh basalts and basaltic andesite that have been dated from other localities. The shapes of these spectra and their deviation from an idealized flat plateau is readily explainable in terms of the combined effects but variable contributions of reactor-induced recoil, low temperature argon loss, and a non-atmospheric “trapped” component (i.e. excess argon). Reactor induced recoil tends to produce anomalously old apparent ages in the low temperature steps and may produce anomalously young ages in the high temperature steps (Fig. 5). Argon loss due to hydration and clay alteration of groundmass glass tends to yield young apparent ages, especially in the lowest temperature steps. Excess argon tends to be trapped in the early-crystallized phenocryst phases (e.g. olivine, plagioclase) and is generally most evident in the highest temperature steps, associated with the lowest apparent K/Ca ratios. All of these complications may be evident to varying degrees in a single sample but do not preclude a fairly precise assessment of the age. In particular, empirical studies by Gans and Bohrson (1998) on basalt samples whose ages are tightly bracketed by sanidine-bearing tuffs have shown that the reasonably flat
The relative stratigraphic position of the samples was unknown. I carefully examined all of the previous samples petrographically, as well as the new samples collected in this study, and grouped them according to their modal composition and textural characteristics, and then used the stratigraphic relations observed in drill core and in the field to assign relative ages to each group. It was clear that the samples fell into three distinct compositional and textural groups (Fig. 6). Where stratigraphic relations were available, the groups had consistent relative ages.

**Age and petrographic groupings**

At the inception of this study, it was known that samples of “Saddleback Basalt” from the Boron area had yielded ages ranging from 21.5 to 18.8 Ma (Kozack, 2000). Many of these earlier analyses had large uncertainties and it was not clear whether this scatter represented real age differences. In most cases, the central to high temperature part of the spectrum corresponding to the gas released from about 800°C to ~1050 °C provides a reliable estimate of the age, even if it does not constitute more than 50% of the gas released or strictly define a statistical plateau. In cases where the individual steps of this segment do not lie within two sigma analytical uncertainty of each other, I assign an uncertainty corresponding to one standard deviation of the selected ages—a somewhat arbitrary assessment but one that has proven conservative in other studies.

**Figure 5:** Representative $^{40}$Ar/$^{39}$Ar age spectra for mafic rock units in the Boron area. All had previously been correlated with Saddleback Basalt, but only the oldest unit is representative of basalts at Saddleback Mountain. Shaded parts of spectra indicate steps used to calculate “plateau” age. Most yield classic reactor-induced recoil type spectra. See text for discussion.

**Figure 6:** Representative photomicrographs of mafic volcanic units in the Boron region. All are plane light and at same scale (horiz bar = 1 mm, FOV width = 5.2 mm). (6A) Sample 04PGMJ-73 of ~20.5 Ma diabase basalt lava from Saddleback Mt. (6B) Sample EBC-10-1158’ from interior of ~19.5 Ma olivine andesite flow in drill core. (6C) Sample EBC 09-880’ from interior of uppermost basaltic andesite lava (~19.2 Ma). (6D) Sample EBC-05-1352’ collected from interior of thick basaltic andesite.

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The groups that were identified included, from oldest to youngest:

a. Coarse-grained (diabasic)

b. Diktytaxitic plag-oliv-cpx basalt lavas

c. Phenocryst-poor oliv andesite lavas

d. Oliv-plag phryic basaltic andesite lavas and compositionally identical sills.

The age groups portrayed in Figure 7 are separated strictly by the petrographic attributes of the samples. The fact that the ages for the groups also cluster provides confirmation that mafic volcanism occurred in three compositionally and temporally distinct pulses, likely coming from different source regions. In addition, each of these compositional groups yielded internally consistent ages, generally distinct from ages for other groups. Samples of the older coarse-grained basalt yield individual ages ranging from ~20.0 to 21.5 Ma, but all within analytical uncertainty of each other, and with an weighted mean age of 20.5 ± 0.3 Ma. Samples of platy olivine andesite lavas from widely dispersed sites gave ages ranging from 19.3 to 19.9 Ma, and likely represent a real range of ages from 19.4 ± 0.2 to 19.8 ± 0.2 Ma. Finally, both the oliv-plag basaltic andesite lavas and compositionally identical underlying sills gave consistent, indistinguishable results with a weighted mean age of 19.2 ± 0.2 Ma. It is worth noting that in every drill hole and surface exposure where relative ages of samples could be determined from stratigraphic relations, the analytically determined ages were always in the correct order (Fig. 2–5).

The youngest (19.2 Ma) age group is best represented by the mafic lavas on the north wall of the Kramer Pit and the large sill complex that underlies the ore body. The age of the tuff within the overlying orebody is 19.28 ± 0.08, analytically indistinguishable from the ~19.2 ± 0.2 Ma age of the underlying sill. The implications of this close temporal and spatial association of borate mineralization and the youngest of the 3 pulses of mafic magmatism are explored further below.

**Mafic volcanic units of the Boron area: a revised nomenclature**

The field and geochronologic data discussed above provides evidence for temporally distinct pulses of mafic volcanism in the Boron area. Petrographic characteristics demonstrate that these groups of volcanic rocks can be distinguished on the basis of their texture and modal composition. Given their distinct ages, stratigraphic positions, and compositions, it is recommended that a revised nomenclature be adopted for the mafic volcanic rocks of the Boron area. Proposed names and a summary of the characteristics for each mafic volcanic rock unit follow:

**Saddleback Basalt (Tsb)— 20.5 ±0.2 Ma**

The type section of the Saddleback Basalt at Saddleback Mountain provides good exposures of the oldest mafic rock unit from the Boron area. It is proposed that this name be retained, but its usage be restricted to the ~ 20.5 Ma coarse-grained olv-plag-cpx basalt that occurs near the base of Miocene sections in the area to the northeast of the mine. This basalt is easily distinguished from all other mafic volcanic rocks in the area and is easily dated.
distinguished by its coarse-grained (diabasic/ophitic) texture with randomly oriented crystals of plagioclase, clinopyroxene, and iddingsitized olivine up to 1 mm in length (Fig. 6A). Upper parts of individual flows may contain interstitial glass and abundant vesicles, but retain the coarse-grained character. This unit commonly displays a conspicuous diktytaxitic texture with coarse plagioclase crystals protruding directly into irregular vapor cavities. The high percentage of mafic minerals and the very low (0.04-0.15) K/Ca ratios derived from the argon analyses suggest that Saddleback Basalt is a primitive tholeite basalt. Estimated ages, K/Ca ratios, and radiogenic yields for samples of Saddleback Basalt that have been dated are given in Table 1.

Table 1. Summary of estimated ages and related data for samples of Saddleback Basalt

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Age (Ma)</th>
<th>± Age</th>
<th>K/Ca</th>
<th>%Rad</th>
</tr>
</thead>
<tbody>
<tr>
<td>USB-1</td>
<td>20.00</td>
<td>0.30</td>
<td>0.05-0.16</td>
<td>27-43</td>
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<tr>
<td>USB-2</td>
<td>20.25</td>
<td>0.20</td>
<td>0.05-0.08</td>
<td>18-41</td>
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<td>IKS-4</td>
<td>20.50</td>
<td>0.20</td>
<td>0.03-0.13</td>
<td>39-90</td>
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<tr>
<td>EBC-10-1316</td>
<td>20.55</td>
<td>0.10</td>
<td>0.038-0.06</td>
<td>57-89</td>
</tr>
<tr>
<td>IKS-6</td>
<td>21.30</td>
<td>0.60</td>
<td>0.02</td>
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<td>04PGMJ-73</td>
<td>20.55</td>
<td>0.15</td>
<td>0.04-0.10</td>
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<td>04PGMJ-72</td>
<td>20.7</td>
<td>0.20</td>
<td>0.04-0.07</td>
<td>60-75</td>
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Stonehouse Andesite (Tsa)—19.4 to 19.7 Ma

Very fine grained, nearly aphyric andesite lavas are exposed in the Murok and Stonehouse Hills, the low hills north and west of the Kramer Pit, and in the deeper parts of drill holes in the east Boron basin area (Fig. 2, 4). It is proposed that this unit be named the Stonehouse Andesite. It typically contains 1–2% small (<200µ) olivine microphenocrysts set in a somewhat trachytic, plagioclase-rich groundmass (Fig. 6B). Several distinct lava flows were identified in some of the drill holes, and this unit is generally interbedded with fine-grained sandstone and siltstone towards the middle of the Miocene sedimentary sections. A likely source area for these lavas is a highly degraded cinder cone of the same material in the low hills ~3 km NE of the Kramer Pit (~2 km NW of Saddleback Mountain). In this area, Stonehouse Andesite lavas rest directly on Saddleback Basalt lavas. The high (0.4-1.5) K/Ca ratios and low percentage of groundmass mafic minerals suggest that the unit is andesitic in composition, despite the presence of olivine phenocrysts. Estimated ages, K/Ca ratios, and radiogenic yields for dated samples of Stonehouse Andesite are given in Table 2.

Boron basaltic andesite lavas and sills (Tbba)—19.2 ±0.2 Ma

A sequence of olivine-plagioclase phryic basaltic andesite lavas and closely associated sills occur within fine-grained lacustrine sections in the Stonehouse Hills, as flows along the north wall of the Kramer Pit, as a thick sill in the bottom of the Kramer Pit, and as flows and sills in many of the drill holes of the east Boron basin. It is proposed that this unit be called the Boron basaltic andesite flow and sill complex, for the excellent exposures in the Boron mine area, and its likely close association with borate mineralization (see below). Rocks of this unit have ~2% olivine and 8–10% plagioclase phenocrysts (up to 1 mm) in a seriate groundmass of plagioclase, olivine, and minor clinopyroxene (Fig. 6C, D). The sills and flows are geochronologically and texturally indistinguishable. In areas where both are present, the sills are emplaced in strata only a short distance beneath correlative lava flows, have abundant vesicles in their upper levels, and produce

Table 2. Summary of estimated ages and related data for samples of Stonehouse Andesite

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Age (Ma)</th>
<th>± Age</th>
<th>K/Ca</th>
<th>%Rad</th>
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<tr>
<td>EBC-13-1713</td>
<td>19.35</td>
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<td>EBC-09-1192</td>
<td>19.40</td>
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<td>EBC-8-792.5</td>
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<td>EBC-09-1286</td>
<td>19.70</td>
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<td>0.45-1.4</td>
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<td>IKS-9</td>
<td>19.4</td>
<td>0.20</td>
<td>0.1-0.22</td>
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<td>IKS-8</td>
<td>19.80</td>
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<td>IKS-5</td>
<td>19.85</td>
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Table 3. Summary of estimated ages and related data for samples of Boron basaltic and andesite sills (Tbba)

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<th>Sample #</th>
<th>Age (Ma)</th>
<th>± Age</th>
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<th>%Rad</th>
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<td>BEX 86-3</td>
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<td>Saddleback</td>
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<td>EBC-05-1352</td>
<td>19.20</td>
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<td>0.24-0.65</td>
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<td>EBC-9-949</td>
<td>19.20</td>
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<td>EBC-11-1104</td>
<td>19.50</td>
<td>1.00</td>
<td>0.20-0.64</td>
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<td>19.2</td>
<td>0.15</td>
<td>0.20-0.60</td>
<td>90-95</td>
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</table>

Table 4. Summary of estimated ages and related data for samples of Boron basaltic andesite flows (Tba)

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Age (Ma)</th>
<th>± Age</th>
<th>K/Ca</th>
<th>%Rad</th>
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<tr>
<td>EBC-9-880</td>
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<td>04PGMJ-77</td>
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<td>04PGMJ-67</td>
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<td>0.1-0.25</td>
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<td>04PGMJ-68</td>
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<td>0.10</td>
<td>.08-0.25</td>
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conspicuous soft sediment deformation in overlying clays. All of these relations suggest that the sills were emplaced into wet lake deposits below the lake bottom. The reasonably high percentage of mafic minerals and moderately low (0.1–0.6) K/Ca ratios suggest that it is a basaltic andesite or andesite. Estimated ages, K/Ca ratios, and radiogenic yields for dated samples of the Boron basaltic andesite sills and lava flows are tabulated below.

**Summary of the eruptive history of the Boron area**

The distinct ages, stratigraphic positions, compositions, and textural characteristics of the mafic volcanic and subvolcanic rocks from the Boron area provide a framework for mapping and correlating these rocks and for understanding the eruptive history of the area. The 20.5 ± 0.2 Ma Saddleback Basalt represents local eruptions of the intrusive basalt early in the Miocene depositional history, roughly coeval with nearby rhyolite to dacite eruptions. It is likely that these eruptions occurred prior to the development of the Boron basin. The ~19.4 to 19.7 Ma Stonehouse Andesite lavas were likely erupted from scattered vents early in the depositional history of the Boron basin, as evidenced by their position near the base of the Miocene lacustrine sections. The Boron basaltic andesite lavas and sills were emplaced after the Boron basin was well established and are commonly intercalated with very fine grained (deep) lacustrine facies deposits. Indeed, the abundance of sills rather than flows may be a consequence of these relatively mafic lavas reaching a point of neutral buoyancy and ponding in low-density lakebeds beneath the lake floor rather than erupting within the lake. Alternatively, basaltic andesite lavas may have flowed into the lake and sunk beneath the upper few meters of unconsolidated mud and then spread out laterally in the subsurface. The first interpretation is favored because in drill holes there are commonly multiple sills, and in the mine, some of the Boron basaltic andesite cuts across the lacustrine bedding.

**Discussion: structure and stratigraphy of the Boron basin**

The close spacing of drill holes in the east Boron basin combined with new insights on the ages and correlation of mafic rocks intersected by these drill holes permit a more detailed assessment of the structure and stratigraphy. Cross sections across the east Boron basin using the drill hole and mine geology help define the overall structure of the area. The NE-oriented cross sections show that the arkose unit generally thickens toward the south or southwest, but is interrupted by at least one intrabasin fault. The east-trending sections generally show a westward component of dip for the Miocene strata as well. Perhaps the most interesting observation to be drawn from these sections is that all of the mafic volcanic rocks are restricted to the northern part of the east Boron basin (Drill holes EBC-6 to 13), which suggests that there may be an important WNW-striking dextral(?) fault that separates the northern part of the basin from the southern part of the basin that includes the Kramer Pit.

Observations from the mine and drill holes east of the mine suggest that much of the fine-grained lacustrine sedimentation in the Boron basin is bracketed between ~19.5 Ma and 18.8 Ma. The precipitation of borate ores at ~19.28 ± 0.08 is closely associated in time and space with the emplacement of the underlying 19.2 ± 0.2 Ma Boron basaltic andesite sill in the mine area, suggesting that this magmatic event may have played an important role in borate mineralization. Regardless of the ultimate source of boron, it seems likely that the emplacement of abundant mafic sills into wet lake sediments just below the lake bottom in what may have been the deepest part of the lake would have promoted a sub-lacustrine hydrothermal system and might have profoundly changed the lake water chemistry and temperature in such a way as to facilitate the precipitation of borates. Some questions that warrant additional investigation include: (a) where were the vents for the hot springs that introduced the boron-rich hydrothermal fluids? (b) What was the ultimate source of all the boron? Might it have been leached from the underplated marine sediments of the Rand Schist? (c) Where were the vents for the Boron basaltic andesite and what role did they play in driving the hydrothermal system that introduced borates to the Boron basin?

The structural and tectonic setting of the Boron basin remains somewhat obscure, but these new results are provocative. It seems likely that the Boron basin represents an extensional basin or half graben that developed in the hanging wall of a N-dipping normal fault (West Boron fault) now exposed on the south side of the Kramer Pit. The fine-grained lacustrine deposits in the southern (deeper) part of this basin fill likely record rapid subsidence associated with slip on the bounding normal fault. The overlying coarse-grained arkosic fanglomerate may record progradation of coarse clastics into the basin, either during continued fault slip, or infilling of the basin after extension ceased. In any case, fine-grained lacustrine deposition, locally-derived basaltic andesite eruptions, and borate mineralization all occurred in a short time interval around 19.2 Ma. It is possible that the bounding fault zone might have served as a conduit for the mineralizing fluids.

**Implications for future borate exploration**

The combined structural, stratigraphic, and geochemical observations from the Boron basin area may provide new ideas for borate exploration in the Mojave Desert. Future exploration efforts might seek to identify and focus on targets that have the following characteristics:

- Stratigraphic sections that include deep lacustrine (claystone) facies
- Stratigraphic sections that accumulated syntectonically in the proximal hanging wall during rapid slip on a basin-bounding normal fault
• Basins where local syndepositional volcanism and intrusion—especially mafic volcanism—was occurring. It might be particularly advantageous to identify sections where sills were being emplaced into the lakebeds while the lake was still active.
• Lacustrine sections that are between ~19.5 and 17.5 Ma, as this appears to be the age of most Boron occurrences and the timing of major regional extension throughout much of the central Mojave Desert.

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Fault-influenced incision in western Grand Canyon, Arizona, U.S.A.

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Abstract—Preliminary interpretation of new and updated incision rates in western Grand Canyon shows the effects of Quaternary faulting, which dampens river incision rates in the footwalls and amplifies them in the hanging walls of normal faults. In the reach between Lava Falls and Diamond Creek in western Grand Canyon, about 178 to 225 river miles downstream from Lees Ferry, the river crosses the neotectonically active Hurricane and Toroweap faults. For this reach, we offer a preliminary analysis of 23 new and updated incision rates determined from new U-Th dating of travertine associated with perched river gravel and previously published 40Ar/39Ar ages on intracanyon basalt flows that overlie river gravel. Results reveal diminished incision rates downstream of both the Hurricane and Toroweap faults, indicating the presence and geometry of hanging-wall anticlines with wavelengths of about 10 km. Upstream of the faults, increased incision rates are interpreted to represent localized footwall uplifts and/or regional block uplift, which could contribute to the overall uplift of the Colorado Plateau.

Past incision work

Bedrock incision rates can be calculated by measuring the height difference between two bedrock strath surfaces and then dividing by the time taken for the river to move from one position to the next. Limitations of the chronometers used to date materials and the context between the dated material and the surface necessitate grouping rates into preferred, maximum, and minimum categories. In Grand Canyon, Quaternary bedrock incision rates of the Colorado River have been estimated in a number of studies (Pederson et al., 2002, 2006, 2013; Karlstrom et al., 2007, 2008; Polyak et al., 2008; Crow et al., 2014, 2018, 2019; Abbott et al., 2015) and currently over 50 individual incision rates are available (Figure 1). These data have been collected to investigate and assess the degree to which bedrock incision has varied through time and space. Temporal variations in incision are expected due to climatic fluctuations (e.g. Pederson et al., 2006) or during rearrangement of river systems (Cook et al., 2009), like during integration of the Colorado River (Karlstrom et al., 2007; e.g. Crow et al., 2021) when a new baselevel was established. However, for at least one million years in Grand Canyon, long-term bedrock incision averaged across glacial cycles has been temporally steady, likely due to recent and ongoing uplift of the Colorado Plateau (Crow et al., 2014; Karlstrom et al., 2007, 2008). In contrast, spatial variation of incision rates has been substantial with temporally steady incision rates of >155 meters per million years (m/m.y.) over the past 1.2 Ma at Lees Ferry at the start of Grand Canyon (Albonico, 2021) (c.f. short term rates of 320 m/m.y. (Pederson et al., 2013)), about 160 m/m.y. over the past ~1 m.y. in eastern Grand Canyon (Crow et al., 2014), and about 90–100 m/m.y. in western Grand Canyon over the past ~4 m.y. (Crow et al., 2014; Polyak et al., 2008). This 60–70 m/m.y. incision rate difference between eastern and western Grand Canyon is interpreted to record several hundred meters of uplift of the Colorado Plateau (Crow et al., 2014; Karlstrom et al., 2007, 2008).

It has also been recognized that over much shorter distances incision rates vary across faults (Karlstrom et al., 2007). Immediately adjacent to faults, footwall rates are equal to hanging wall rates plus the fault slip rates. However, at distances of about 10 km from the faults, rates often return to far-field rates due to fault-related flexures and diminishing offset (e.g. Walk et al., 2019). At broad spatial scales, faults may generally coincide with the boundaries between blocks being uplifted at different rates, but the goal of this contribution and on-going work is to parse local fault-related effects (fault slip and flexure) from broader mantle-driven epeirogenic uplift effects. These relations, and specifically the nature and geometry of fault-related flexures, will be explored here in the context of the new and updated incision rate data.

Grand Canyon’s lava dams

Many of the updated incision rates detailed here are related to Grand Canyon’s basaltic lava dams, which cap and preserve perched river gravels. Crow et al. (2015) reported the results of a 40Ar/39Ar dating campaign...
coupled with geochemistry, paleomagnetic analysis, and field relationships to better understand these dams, building on the work of Hamblin (1994), Fenton et al. (2002, 2004, 2006), and others. The new chronology suggests at least 17 damming events from 830 to 100 ka. Some flows erupted directly into the canyon and others had vents on the rim that sent lava flows cascading into the canyon. In both cases, lava pooled at the canyon bottom before flowing downstream, in some cases more than 135 km. The resulting lava dams were removed by the Colorado River, likely in hundreds of years or less, leaving a series of isolated remnants, some of which preserved perched gravels and can be used to calculate incision rates. Incision rates based on early dating of those dam remnants have been reported (Pederson et al., 2002; Karlstrom et al., 2007, 2008), but some rates were based on incorrect assumptions about the ages of flows and other now-dated remnants have never been used to calculate incision rates. Those new and updated rates are investigated here for the first time.

**Effect of Plio-Pleistocene faulting on incision**

Tectonics have long been recognized as having influenced Colorado River evolution and deformed its deposits (Howard and Bohannon, 2001; Pederson et al., 2002; Karlstrom et al., 2007; e.g. Howard et al., 2015; Seixas et al., 2015). Major normal faults crossing the course of the Colorado River in or near western Grand Canyon include: the Hurricane and Toroweap faults in western Grand Canyon (Karlstrom et al., 2007) and the Wheeler, Fortification, and Detrital Valley faults in the Lake Mead area (e.g. Crow et al., 2019). Pederson et al. (2002) and Karlstrom et al. (2007, 2008) proposed the concept of fault-dampened incision where the difference in incision rates upstream and downstream of active normal faults was equal to the fault throw. However, incision reflects a combination of fault-related flexures (Howard and Bohannon, 2001; Hanks and Blair, 2003; Karlstrom et al., 2007; Seixas et al., 2015) and epeirogenic mantle-driven uplift (Crow et al., 2014). Walk et al. (2019) examined differential incision of the Virgin River as it crossed the Hurricane normal fault system in Utah and applied the concept that incision rate differences across and at least 5 km from faults was a proxy for Colorado Plateau uplift but that rate differences immediately across the fault also record formation of hanging wall anticlines. This paper aims to parse local from regional fault-influenced differential incision in Grand Canyon based on new incision rate data.

**New and revised incision rates**

Perched Colorado River gravels throughout Grand Canyon have been used to quantify incision. Figures 1 and 2 show the vast majority of incision rates calculated in Grand Canyon from dated deposits or surfaces related to the Colorado River. Twenty-three of those incision rates are newly reported or revised by this study in a preliminary fashion. New geochronology includes five U-Th analyses on travertine (four of those gave robust ages), yielding four new incision rates. Five other new incision rates are based on previously published 40Ar/39Ar ages from lava dam remnants which overlie Colorado River gravel (Crow et al., 2015). Additional incision rates, from 40Ar/39Ar ages reported in Crow et al. (2015), revise previously published rates (Pederson et al., 2002; Karlstrom et al., 2007). Larger differences (up to 120 % change) in incision rate are due to inclusion of additional 40Ar/39Ar analyses (Crow et al., 2015) from samples

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Some rates considered to be maxima with limited incision significance by the original authors have been excluded for brevity.
collected at sites originally detailed by Pederson et al. (2002) and/or Karlstrom et al. (2007). In some cases, the early preliminary incision rates were based on assumed ages from dating of flows thought to be correlative; not all of these correlations have been supported by additional study. In other cases, additional analyses from the same samples have refined the preferred age of the lava flow remnant and the updated weighted mean ages were used to calculate revised incision rates.

Our preferred basalt-constrained incision rates range from ~ 50 to 175 m/m.y.—a result based on dating of 100 to 830 ka lava dam remnants overlying Colorado River gravel. Maximum basalt-constrained rates range from 21 to 297 m/m.y. Four new incision rates are based on U-Th dating of travertine infilling and veins in Colorado River gravel at river mile (RM) 2190 and 273. At RM 273, a laminated travertine vein containing river sand and colluvium cuts bedrock and Colorado River gravel. The strath is approximately 125 m above pre-Lake-Mead river level and the cross-cutting vein was dated at 480 ka yielding a maximum incision rate of 308 m/m.y. We also dated a lower, ca. 200 ka travertine drape not associated with Colorado River gravel in this area yielding a maximum rate of 600 m/m.y. Because these deposits only constrain the Colorado River to have been below them at the time of travertine deposition and because of the extreme rates, they likely have little to no incision significance. U-Th dating of travertine infilling associated with Colorado River gravel at RM 190 yielded maximum incision rates of a little over 190 m/m.y., which provides important new incision constraints in the footwall of the Hurricane fault.

**Discussion and conclusions**

Incision rate variations occur spatially throughout Grand Canyon on a variety of scales (Figure 2). Short wavelength variations in incision rate appear to occur exclusively in association with faults. The most pronounced variations are associated with the Hurricane and Toroweap faults. The new and updated incision rates in proximity to the Hurricane fault show that incision rates on the downstream side decrease from about 90 m/m.y. to 50 m/m.y. toward the fault. About 80% of the decrease in incision rates in the hanging wall of the Hurricane fault occurs in the 10 km immediately downstream of the fault. This is interpreted to represent a hanging wall anticline (Hamblin, 1965, 1984; Hamblin et al., 1981; Karlstrom et al., 2007) with a comparable wavelength, which damps incision more effectively closer to the fault where the displacement is greatest.

Across the Hurricane fault, in the footwall, maximum incision rates calculated over the last 350-570 ka are about 190 m/m.y., and the difference in rate across the fault is slightly higher than but consistent with and explainable by the fault slip rate of about 75-100 m/m.y. (Fenton et al., 2001; Karlstrom et al., 2007). An exception to this relation is a new incision rate calculated from the dating of an 830-ka basalt flow at RM 188.8. The older than expected mean age of 829 ± 9 ka was verified by duplicate analyses on two samples using varied analytical setups (Crow et al., 2015). The samples were collected from a peperite outcrop where glassy basalt is mixed with river sand. It yields an incision rate of 73 m/m.y., less than predicted for its location in the footwall of the main Hurricane fault strand. Two possible explanations for the low incision rate include undetected 39Ar recoil and/or Quaternary slip on fault splays farther east from the main strand of the Hurricane fault. Although the step-heated age spectra for all four replicate ages don’t show obvious evidence for recoil, the lowest temperature steps were not analyzed because other samples produced very large signals with low radiogenic yields at low temperature, which affected the blanks for subsequent steps (Karlstrom et al., 2007; Crow et al., 2015). Alternatively, it is plausible that the outcrop is in the hanging wall of a Hurricane fault splay with Quaternary slip. Billingsley and Wellmeyer (2003)

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2 Measured downstream from Lees Ferry (Stevens, 1983)
mapped a fault, which we suspect may have Quaternary offset, as being covered by remnants of the Whitmore and Upper Gray Ledge flows, which have been dated to between 250 and 100 ka (Crow et al., 2015), respectively. If this fault was active between 830 and 250 ka, the lower-than-expected incision rate could be a consequence of fault dampering.

Upstream from the Hurricane fault and its splays, incision rates again decrease in the upstream direction towards the Toroweap fault, in the ~12 km wide block between the two faults. Rates decrease from less than 190 m/m.y. to about 120 m/m.y. halfway across the block. The ca. 485 ka “Toroweap A” basalt flow is cut by the Toroweap fault and gives incision rates of 65 and 162 m/m.y. in the immediate hanging wall and footwall, respectively. The difference is explainable by the 100 m/m.y. slip rate of the Toroweap fault (Fenton et al., 2001; Karlstrom et al., 2007). Generally, incision rate variations in the block seem to be consistent with folds that amplify incision in the footwall of the Hurricane and folds that diminish incision in the hanging wall of the Toroweap fault. However, a newly determined maximum rate of 38 m/m.y. in the center of the block between the Hurricane and Toroweap faults is an apparent exception to the decreasing rates. This incision rate comes from the ca. 490 Ma 183.4-mile basalt flow, whose base is below river level. However, this remnant is inset into basaltic cinders and likely fell to its current location (Crow et al., 2015), which could explain why the resulting incision rate is much lower than expected.

Immediately upstream from the Toroweap fault, two incision rates have been calculated, one of 162 m/m.y. as discussed above and another about 3 km further upstream of 144 m/m.y. This suggests potentially another short-wavelength flexure in the footwall of the Toroweap fault. Given the available information it is also conceivable that the entire block east of (upstream of) the Toroweap fault is being uplifted semi-uniformly (Karlstrom et al., 2007, 2008) as the incision rates in the footwall of the Toroweap are roughly equivalent to those in eastern Grand Canyon (about 160 m/m.y., Crow et al., 2014). Incision rates in the Elves Chasm (RM 116) and Surprise Valley (RM 135) areas between the Toroweap fault and eastern Grand Canyon are variable—determined to be about 100 m/m.y. at Elves Chasm (Crow et al., 2014) and ≥ 138 m/m.y. at Surprise Valley (Robertson et al., 2021) USA. The latter is similar to rates in eastern Grand Canyon and in the immediate footwall of the Toroweap fault and the former is similar to rates in western Grand Canyon. Robertson et al. (2021) suggested that the lower incision rate at Elves Chasm could potentially be related to dampering by downstream landslides. If not locally affected by landslides, the Elves Chasm rate would support the idea of localized uplift in the footwall of the Toroweap fault as opposed to uniform block uplift, whereas the Surprise Valley rate is more consistent with semi-uniform block uplift.

Both localized and broad uplift in footwalls of normal faults are expected and are not mutually exclusive. Elastic modeling indicates not only the presence of localized (across tens of km) uplift in the footwalls of normal faults following fault rupture, but also suggests that footwall deformation can indicate the geometry of faults at depth (Resor and Pollard, 2012). Geodetic observations also confirm areas of broad (~100 km) uplift parallel to Basin and Range normal faults decades after large rupture events, likely indicative of aseismic uplift due to mantle flow according to modeling studies (Thompson and Parsons, 2017).

The new incision rate data in Grand Canyon indicate localized incision rate variations consistent with the formation of hanging-wall anticlines associated with both the Toroweap and Hurricane faults as well as elastic fault rupture. These folds appear to have a wavelength a little over 10 km, which is also crudely seen in the folding of much older strata (Billingsley and Wellmeyer, 2003). In addition to localized footwall uplift, there are also some incision rate data indicative of broad uplift upstream from the Toroweap fault, which could contribute to the uplift of Colorado Plateau margins. Additional incision rate measurements in central Grand Canyon could be helpful in further testing these ideas.

References cited


Recently discovered warm springs at Obsidian Butte, Imperial County, California

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ABSTRACT—Obsidian Butte is one of the Salton Buttes, which are five Holocene rhyolitic volcanic domes associated with a relatively shallow (~4 km) magma body in the Salton Sea Geothermal System. Late in 2020, several small springs were discovered flowing into the sea along the shore north of Obsidian Butte. The springs span about 150 m along the shore and were recently exposed as the sea level has dropped. They have water temperatures as high as 27°C and form small channels that cross mud flats and bars before entering the sea. Over the period of one year the outflow channels from the springs have changed as the sea level has further dropped, as have the flow rates and temperatures of some of the springs. The combined flow of several of the springs in the main outflow channel is significant (~400 liters/min) with a velocity of nearly 30 cm/s. Thermal imagery showed that water in spring channels with significant flow retained elevated temperatures as it traversed the mud flats. In view of the many geothermal features in the vicinity, the springs’ slightly elevated temperatures may be geothermal in origin. We note, however, that spring temperature may be influenced by the region’s mean annual temperature of 25°C, which controls the near-surface temperature of the rock and sediment through which the waters flow before emerging.

1. Introduction
The Salton Buttes are five rhyolitic volcanic domes associated with the Salton Sea Geothermal System (SSGS) located near the southeast shore of the Salton Sea in Imperial County, California. The SSGS lies in the Salton Trough, a pull-apart basin located at a tectonic stepover between the San Andreas and Imperial faults (Gurbuz, 2010) and has a high geothermal gradient anomaly that is related to the numerous geothermal plants of the Salton Sea Geothermal Field (SSGF) (Hulen et al., 2002). The Salton Buttes, from north to south, are Mullet Island, north and south Red Hill, Rock Hill, and Obsidian Butte. They collectively form a roughly north-northeast trending line about 7 km long. The buttes were thought to have formed during the late Pleistocene (Muffler and White, 1969), but recent work suggests eruptions as recent as 2.45 ka for Red Hill and Obsidian Butte (Schmitt et al., 2013; Wright et al., 2015). Hulen et al. (2003) give a detailed description of the subsurface geology of the Obsidian Butte sector of the SSGF based on analysis of several thousand feet of borehole rock samples. Obsidian Butte is the surface expression of the Obsidian Butte flow-dome-tuff complex (Robinson et al., 1976) which consists of glassy to stony, sparsely porphyritic, flow-banded and pumiceous to perlitic rhyolite. It is underlain by the Pleistocene to Holocene Brawley Formation and uppermost Borrego Formation. The former contains a several-hundred-meter-thick evaporitic anhydrate-bearing mudstone that forms a natural cap to the geothermal system. In the conceptual model of Hulen et al. (2003), the source of the geothermal energy for the Obsidian Butte sector of the SSGF is a still-cooling granitic pluton 2 km in diameter and 3.5 km below the surface. The production fluids are circulating hypersaline brines with temperatures generally greater than 290°C. At depth, the interaction of the brines with the wall rocks has led to hydrothermal alteration and development of epidote, actinolite and clinopyroxene.

Obsidian Butte is located in the Brawley seismic zone and from 28 August to September 2, 2005 a swarm of more than 550 earthquakes with magnitude 1.0 or greater occurred near Obsidian Butte (https://www.cisn.org/special/evt.05.09.01/). Eight events were greater than magnitude 4.0 with the largest being 5.1. Its epicenter was 9.7 km deep and was located 1.6 km south of the butte and 2.3 km south of the wells. Analysis of earthquake data indicated that seismic faulting was restricted to a depth of about 4–6 km along a 10 km-long NE striking near-vertical fault segment. In a cross section based on drill core data, Hulen et al. (2003) show a high angle fault near the zone defined by the earthquake epicenters. Surface deformation was observed by two nearby GPS stations, as were some surface breaks in the field (Lohman and McGuire, 2007).

Obvious surface expressions of geothermal activity in the SSGS occur at fumarole fields 1 km SE of Mullet Island (Lynch et al., 2013; Adams et al., 2017) and at the intersections of Davis and Schrimpf Roads, 4 km SE of Mullet Island (Onderdonk et al., 2011), with the former being currently the most active. The fumarole field 1 km SE of Mullet Island consists of vents with boiling water and mud and small gryphons that are emitting ammonia (Tratt et al., 2016), which produced
Table 1. Dates of observations and measurements.

<table>
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<th>Date</th>
<th>Sea Level (m)</th>
<th>Measurements</th>
<th>Observations/Comments</th>
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<td>Initial discovery of springs</td>
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<td>Drone (DJI Mavic Pro) survey (DKL)</td>
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<td>02/01/2022</td>
<td>-72.838</td>
<td>Limited photo survey, flow rate measurement</td>
<td>Algae in spring discharge very pronounced</td>
</tr>
</tbody>
</table>

Lynch and Adams (2014) did not detect any thermal anomalies on Obsidian Butte (elevation -39.62 m) when they surveyed it in January of 2014, but they concentrated on rock outcrops similar to those that had thermal vents on Red Hill and did not survey the shoreline (sea level = -70.90 m). Since that time, the Salton Sea level has dropped about 1.83 m (Figure 1), recently exposing several areas (OB1-OB6) discharging water (springs). Table 2 gives their locations. Figure 2 shows the locations of the springs superimposed on a 2019 Google Earth image. Figure 3 shows a drone image of the shoreline on January 17, 2021.

In February 2021, a survey of the shoreline surrounding Obsidian Butte was made and the only area detected with flowing water spanned about 150 m between the points OB1 and OB2. About 10 easily recognizable springs emerged from loose barnacle shells and mud between cobbles. Upslope from the springs onto a mudflat and into the Salton Sea. There was no evidence of an upslope channel that supplied the water. At their sources they were found to be warm to the touch and an uncalibrated thermometer suggested the water temperature was 27°C. This initiated a more detailed study (Table 1).

2. Observations

Lynch and Adams (2014) did not detect any thermal anomalies on Obsidian Butte (elevation -39.62 m) when they surveyed it in January of 2014, but they concentrated on rock outcrops similar to those that had thermal vents on Red Hill and did not survey the shoreline (sea level = -70.90 m). Since that time, the Salton Sea level has dropped about 1.83 m (Figure 1), recently exposing several areas (OB1-OB6) discharging water (springs). Table 2 gives their locations. Figure 2 shows the locations of the springs superimposed on a 2019 Google Earth image. Figure 3 shows a drone image of the shoreline on January 17, 2021.

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Figure 1. Variation in the Salton Sea level (m) from 12/1/2014 to 2/1/2022. From: (waterdata.usgs.gov/nwis/dv?referred_module=sw&site_no=10254005). The sea level fluctuations are periodic because of seasonal variations in temperature, rainfall, and agricultural runoff.
there were no recognizable drainage channels that could supply the water. About 25 m south of OB3-OB5 there were outcrops of obsidian and rhyolite to 6 m high that would serve as a barrier to drainage from the butte. At the points of emergence, the springs formed small channels 2–20 cm wide and 1–5 cm deep that merged to form larger shallow channels 10–40 cm wide and 2–10 cm deep that flowed north across sand bars consisting of barnacle shells or mud flats to the open sea (Figure 3–5). There was common enhanced vegetation growth at the source of the springs. Scattered intermittent small bubbles were observed emanating from the spring sources, as well as locations up to several meters from the points of emergence. The temperatures of the five major springs (OB1-OB5) at their sources was 27°C. Water temperatures were measured with a Taylor digital thermometer with a two-meter-long cable probe. The Salton Sea water temperature, measured 40 cm from the shoreline at BL2 away from the springs, was 14°C on 2/20/2021. The temperatures of the Obsidian Butte springs are considerably less than the mudpots 1km SE of Mullet Island (to 100°C) and the Davis-Schrimpf gryphons (40°–70°C) and less than the temperature of the thermal vents on Red Hill (38°C). The total dissolved solids (TDS) content of water collected from OB5 on 2/21/2021 was 1.3%, which compares with 7.4% for Salton Sea water taken from location BL1 in November 2021. The residue was analyzed by X-ray diffraction and was found to be primarily halite with minor anhydrite and a trace of hexahydrate.

When the area was visited in April 2021 the sea level was slightly higher than in February and there appeared to be more water on the mud flats between the sources of the springs and the open sea. This made the outflow channels less distinct. The temperatures of OB1–OB5 were still 27°C with the exception of OB2, which was 17°C. The open sea water was 20°C.

On November 27, 2021, the level of the Salton Sea had dropped 3.3 cm compared with the last visit in April and the appearance of the shoreline and the configuration of the springs had changed

Table 2. GPS coordinates of discharge sites (OBs) and baseline (BL) Salton Sea measurement sites.

<table>
<thead>
<tr>
<th>Sites</th>
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<th>Longitude</th>
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<tbody>
<tr>
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<td>W 115° 38.46′</td>
</tr>
<tr>
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<tr>
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</tr>
<tr>
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<td>W 115° 38.57′</td>
</tr>
<tr>
<td>BL2</td>
<td>N 33° 10.48′</td>
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Figure 2. Locations of springs (2/2021) superimposed on 2019 Google Earth image. Locations OB1 and OB2 represented the western- and eastern-most locations, respectively (1/2021 to 4/2021). OB6 was subsequently exposed and OB2 was inactive on 11/ 2021. BL1 and BL2 are locations for Salton Sea baseline measurements. OBS-01 and IID-08 are capped geothermal wells. See Table 2 for GPS coordinates.
dramatically. At this time, images of the shoreline were taken with a point and shoot camera (Olympus TG-3, 12s timer delay) mounted on a 7-m-long telescoping pole (Pole Cam), in order to provide an elevated panoramic perspective when the drone was unavailable. Earlier in 2021, the exact shoreline in this area was difficult to firmly establish. There appeared to be a shallow sand bar and a partially water-covered mud flat separating the springs from open water (Figure 4). It was unclear if the water on the mud flat was from the Salton Sea or an accumulation from the numerous springs. In November 2021, and subsequent visits, the shoreline was much better defined (Figure 6). Proceeding north from the area between OB3 and OB5, there were: 10 m of 15-30 cm-deep loose barnacle shells, a 1m-wide outflow channel with rapidly flowing water, 15 m of salt encrusted firm mud, 15 m of moist mud and 15 m of wet mud that was too soft to walk on.

A new area of flowing small springs (OB6 in Figures 2, 6), that had previously been under water, was also discovered further west from OB1 which earlier had been the western-most. This consisted of 3–4 small (1–3 cm wide) flows emanating from the mud between cobbles within 2 m of the shoreline. Several of the areas previously identified with well-defined springs of flowing water (OB2, OB3) no longer showed evidence of flow but did contain pooled water. When the ambient air temperature was 7°C, the temperature of OB5 and OB6 was 27°C while OB1 = 21°C, OB2 = 7°C, OB3 = 23°C and OB4 = 11°C. The total dissolved solids content of OB5 at the source was 0.34%, which was considerably lower than that measured on February 2021 when the sea level was higher.

The channel leading from OB5 had undergone significant changes. Early in 2021 it wandered in a northerly direction (Figure 3) but in November 2021, and in subsequent visits, it travelled for ~40 m west along the old shoreline in a well-defined channel before making a right-angle bend north to open water (Figures 6, 8). Figure 7 shows OB5 on December 2020 and December 2021. On the later date, OB5 showed two main areas emanating flowing water. Much of the area surrounding the springs was now covered in deep (> 15 cm) loose barnacle shells which covered most of the larger cobbles. In OB5’s larger emergence point, the loose barnacle shells could be seen jostling/levitating in the...
flow. There was also noteworthy vegetation growth. This was typical of the areas around the other earlier identified springs and in many cases it was not possible to precisely locate the earlier locations because points of reference (cobbles) had been covered by barnacle shells or obscured by vegetation (tamarisk and a form of rush). In the swiftly flowing OB5 outflow channel, there was abundant aquatic insect activity (water boatmen, water striders) along with a school of 3 cm-long fish that resembled pupfish.

Measurements of the flow rate from OB5 were made at two locations in November 2021; at the larger of its two emergence points (Figure 7) and in the fast moving outflow channel (Figure 8). Bamboo stakes were placed along the side of the channels at 25 cm intervals and the cross-sectional areas of the flow were measured at three locations. A video camera was used to record the movement of a Styrofoam packing peanut carried by the flow and analysis of the video was used to determine the velocity. At the emergence point, the mean cross-sectional area for the first 0.75 m was $542\pm57$ cm$^2$ and the velocity was $3.2$ cm/s which represents a mean flow rate of $104\pm11$ liters per minute. Further along the channel (Figure 8) the water temperature was $25^\circ C$ and the mean cross-sectional area of the flow was $219.4 \pm 12$ cm$^2$ with a velocity of $29.4$ cm/s which represents a mean flow rate of $387\pm21$ liters/min. This is significantly greater than the flow rate at the source of OB5 but there were several other small flowing sources to the west of OB5 that contributed to the flow along with any combined diffuse seepage. The flow rate at the location in Figure 8 was remeasured on February 1, 2022. At that time, the mean cross-sectional area of the flow was $229.37 \pm 42.6$ cm$^2$ with a velocity of $33.3$ cm/s which represents a mean flow rate of $458.3 \pm 85.1$ liters/min. While this was greater than in November, there was greater uncertainty because of variations in the cross-sectional area. It was not possible to remeasure the flow rate at the emergence point of OB5 (Figure 7) because by February 2022 the surface had become nearly completely covered with filamentous algae. In general, algae was much more common in the spring fed waters on February 1, 2022 and in many places formed mats completely covering the surface.

To better understand the springs, on February 21, 2021, April 10, 2021 and December 29, 2021, infrared surveys of the area were made with an Agema Thermovision 570 camera (wavelengths ~8–13 micrometers) using 12° and 45° lenses. To have maximum thermal contrast between the springs and surroundings, the surveys were conducted at dawn. This also allowed visible images to be taken for reference. The ambient air temperature during the surveys was 5°C–7°C. The thermal images covered only a limited field of view, so overlapping images were taken and mosaics of the area were constructed. Thermal and visible images taken of OB1 on 2/20/2021 are presented in Figure 9. They clearly show that the water issuing from the spring remains confined to relatively narrow channels as it crosses the bars and mud flats and is significantly warmer than the surrounding waters. Rock outcrops 25 m south of the springs were imaged with the infrared camera but no thermal anomalies were observed.
survey of the outcrops was not comprehensive. In general, the channels that were well defined in February were less evident in April. The thermal images had less contrast because the temperature differences between the ambient air (13°C), the sea water (20°C) and the spring water 27°C were smaller. Thermal and visible images taken of OB5 and OB1 on 12/29/2021 are presented in Figures 10 and 11, respectively. OB5 is warmest (27°C) at its dual sources but the outflow channel remains well above ambient (23°C vs 7°C) based on the thermal images and physical measurements. At OB1, which no longer showed obvious flow, elevated temperatures (21°C) were generally limited to near the source and the majority of the channels through the mud flats were at ambient.

3. Discussion

Based on the most recent eruption date of 2.45 ka for the Salton Buttes (Schmitt et al., 2013), the U. S. Geological Survey has classified them as active volcanoes (https://www.usgs.gov/volcanoes/salton-buttes). Obsidian Butte potentially could be considered as a geothermal source for the elevated temperature of the springs.

Two capped geothermal test wells (IID-8 and OBS-01) drilled by Cal Energy Operating Corporation are located 200 m south of, and upslope from, the springs (Figures 2 and 12). Observation well IID 8 was drilled in 1990 to a depth of 1988 m, with a bottom-hole temperature of 288°C, had TDS of 22% and showed a pressure drop of 55.2 kPa per year (Woods, 2002). OBS-01 was a shallow (30–90 m) temperature-gradient hole drilled between 1984 and 1986 (Woods, 2002). The wells are listed in the California Department of Conservation Well Finder well records (https://maps.conservation.ca.gov/doggr/wellfinder/#openModal/-115.63329/33.16986/16) but there was no record of depth to the water table. In the California Department of Water Resources Database (https://cadwr.app.box.com/v/WellCompletionReports/folder/77225216507), the nearest well with a reported water table depth was located about 1.5 km away at the SW corner of Boyle and McNerney Roads at an elevation of -68.88 m. The well was drilled in 1989 and standing water was encountered at 2.13–2.90 m below the surface. At that time the Salton Sea level was -70.10 m. It is unclear if this can be related to current conditions if the water table is linked to the sea level.

However, there may be a nongeothermal explanation for the elevated temperatures of the springs. It is possible that the rock through which the springs are flowing is warmer than the sea. If so, the springs should show thermal contrast when the ambient temperature is low, such as before sunrise during the winter or early spring when many of our observations were made. The temperature of spring water is controlled in large part by the discharge rate (Manga and Kirchner, 2004). If the groundwater velocity is high, the heat added by geothermal warming is diluted into a large volume of water, thereby cooling the spring temperature. In contrast, if the discharge velocity is low, the subsurface temperature gradient is little affected by the ground water flow and the discharge temperature will be close to the mean annual surface temperature (Manga and Kirchner, 2004).
For Niland, which is 15 km from Obsidian Butte, the mean daily temperature ranges from 15°C to 35°C (https://www.weather-us.com/en/california-usa/niland-climate#temperature) so that the mean annual temperature is on the order of 25°C, similar to the observed temperature of the springs.

Meinzer (1923) divided non-thermal springs into two categories: 1) those with temperatures approximating the mean annual air temperature and 2) cold springs, with temperatures colder than the mean annual air temperature. Nathenson et al. (2003) proposed the term “slightly-thermal” springs for those that do not meet the requirement for thermal springs (temperature > 10°C above mean annual air temperature) but which have dissolved constituents characteristic of thermal waters. This description possibly fits the springs on the north side of Obsidian Butte. However, it may be difficult to distinguish the dissolved constituents as being from deep geothermal brines, Salton Sea water, agricultural runoff, or some combination of all three. The most recently measured TDS content of the main OB5 spring (0.34%) is considerably less than that of the deep brines (22%) or Salton Sea water (7.4%). However, Williams and McKibben (1989) report that the salinities of geothermal brines in the SSGF are distinctly bimodal, with one population with TDS > 20% and another with low TDS. In the latter, they report TDS contents of about 1.5% for some mudpots. TDS contents similar to OB5 have been measured (0.3%) from some boiling pools (at F1) in the fumarole field 1 km SE of Mullet Island (Adams and Lynch, 2014).

4. Summary and conclusions

This paper describes the discovery of small springs along the shore of the Salton Sea north of Obsidian Butte that were recently exposed as the level of the Salton Sea dropped. The springs have temperatures as high as 27°C. Over the period of one year, the outflow channels from the springs have changed as the sea level has further dropped, as have the flow rates and temperatures of some of the springs. The combined flow rate of several of the springs is significant (~400 liters/min) with the main outflow channel having a velocity of nearly 30 cm/s. There are a number of springs in the Salton Buttes area that range in temperature from boiling, at the fumarole fields 1 km SE of Mullet Island (Lynch et al., 2013; Adams et al., 2017), to much lower temperature mound springs at the intersection of Davis Road and State Route 111 (Lynch and Hudnut, 2008; Adams and Lynch, 2014), including an enigmatic moving spring (Lynch and Deane, 2019; Deane and Lynch, 2020), previously known as W9 (Lynch and Hudnut, 2008). The temperature of the latter has ranged from 22°C–24°C (D. Lynch, pers. comm.) with a TDS content of 1.8% measured in October 2021. Even though their temperatures are significantly different, the springs all appear to be partly driven by rising carbon dioxide produced by the decomposition of carbonate sediments by the magma body underlying the Salton Sea Geothermal System (Muffler and White, 1969). The temperatures and TDSs of the springs at Obsidian Butte are similar to those of W9 but show only traces of gas escape. The temperature of the springs may not be completely geothermal in nature since the mean annual temperature of the area is so high and influences the temperature of the surface rocks through which the springs flow, creating thermal contrast during colder times of the year. They may possibly be described as slightly-thermal springs.
References


Joshua tree mortality and recovery after the Dome Fire, Mojave National Preserve

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ABSTRACT—The Dome Fire of August 2020 destroyed more Joshua trees than any previous wildfire and with the lowest survival rate yet recorded. High mortality was associated with the eastern Joshua tree’s (*Yucca jaegeriana*) short, compact stature. In the decade prior to the fire, eight successive years of below average precipitation, followed by 12 months of drought, were considered important contributing factors to the high death rate.

Introduction

On 15 August 2020, a lightning strike started a massive wildfire on Cima Dome in California’s Mojave National Preserve (NPS, 2020). By the time of containment, four days later, 17,512 hectares had burned (Boxall, 2020; Figure 1). Though the Dome Fire was not the most expansive fire to have occurred within a Joshua tree woodland, it was the most destructive in history. An estimated 1.3 million eastern Joshua trees (*Yucca jaegeriana*) were destroyed in nine days (NPS, 2020; Figure 2). Included within the burned area, was a one-hectare Joshua tree study site I had established in 1987 and monitored annually through 2016 (Cornett, 2018; study site coordinates are 35°19’4.4"N, 115°32’51.4"W).

When I inspected the study site on 1 September 2020, only four Joshua trees were undamaged. Each had established along a barren sliver of exposed bedrock that was several meters from the flames. The remaining 128 Joshua trees within the study site had either brown leaf rosettes from the fire’s heat, were erect but charred and blackened, or remained as piles of ash. Large trees that had died but remained standing prior to the fire, collapsed and were left as ash ghosts on the ground (Figure 3).

Figure 1. Extent of Dome Fire (in red). Yellow asterisk is location of 1-ha study site.

Figure 2. Repeat photographs of Cima Dome Joshua tree study site, Mojave National Preserve. Top photograph taken in 1982. Bottom photograph taken on 5 September 2020.
Background

The western Joshua tree, *Yucca brevifolia*, has several characteristics that enable it to survive wildfires. Tall, mature trees, without low (< 2 m) branches usually lack dead leaves on trunks, leaves that would otherwise provide an easy fuel path to the crown. These same trees also eliminate shrubs and some grass species at their base due to competition for moisture and light (Minnich, 1995; Cornett, personal observations). As a result, a fire may have insufficient heat, lowered flames, or move through an area too fast to impact crowns of large Joshua trees (Figure 4). A second characteristic is the production of new leaf rosettes from charred branches after the crown has burned (Figure 5). In this rare response, fire is confined to the crown and the trunk is unscathed. I have only witnessed this in one portion of the Juniper Complex Fire in Joshua Tree National Park (Cornett, 1994; Loik et al., 2000; DeFalco et al., 2010). No regrowth responses have been described for the eastern Joshua tree, *Y. jaegeriana*, the species found on Cima Dome. The latter species is described as having a more compact crown and a shorter trunk (Lenz, 2007; McKelvey, 1938; Rowland, 1978). If accurate, a shorter trunk would place the crown of an eastern Joshua tree closer to the flames of a wildfire and presumably make it more likely to be damaged or destroyed.

In this preliminary study of fire impacts, I hypothesized mortality would be greater for *Y. jaegeriana* on Cima Dome than for *Y. brevifolia* populations in and around Joshua Tree National Park. Higher mortality would be due to (1) the shorter trunk and denser crown of *Y. jaegeriana* as well as (2) possible effects of increased temperature and drought associated with climate change.

Methods

I collected data during the week of 16 December 2021, 16 months after the fire. Trees were counted and included in the study if exposed to flames or heat (as determined by blackening of branches and trunks or browning of leaf rosettes). The occasional tree within the burn area that
showed no signs of fire or heat impacts was not included. Damaged trees were considered to have survived if one or more sprouts emerged from the charred trunk base (root crown sprouts) or if one or more leaf rosettes in the canopy were partially green. Three decades prior to the fire, individual Joshua trees within the study site were marked with numbered aluminum tags affixed to each trunk. Additional survival data were collected from forty, 100 x 10 m-wide belt transects extending beyond west, east, and north study site boundaries at ten-meter intervals for a combined area of five hectares. (Boulder outcrops and unburned areas prevented contiguous data collection to the south.) Numbers and statuses of Joshua trees in the belt transects were included with those of trees within the 1-ha study site for a total of 551 trees evaluated. The above data were compared with three previous studies that examined fire survival rates of the western Joshua tree, *Y. brevifolia*, based upon emergence of root crown sprouts (Table 1; Cornett, 1994; Loik et al., 2000; DeFalco et al., 2010). An independent two sample t-test was used to determine if differences in means was significant. Searches for seedlings were conducted but I found none within the study site or belt transects.

To compare differences in structure of the two species of Joshua tree, I used available data from nine study sites (Table 2). Four sites contained *Y. jaegeriana* and five contained *Y. brevifolia*. The percentage of trees with heights exceeding 3.5 m and percentage of trees with leaf rosettes below 3.5 m were calculated for each site and compared. A chi-square test was used to determine if the variables for the two species were significantly different (Table 2). Correlation coefficients were calculated for rate of survivorship and precipitation, both before and after the Dome Fire, in search of meaningful relationships.
I obtained precipitation and temperature data for the study site and comparison sites from the PRISM Climate Mapping Program, Oregon State University, at the 4 km grid resolution level. Regional summarized precipitation and temperature data were obtained from the Western Regional Climate Center (data accessed at https://raws.dri.edu/ on 24 December 2021).

### Results

Eastern Joshua trees are thought to be shorter and more compact than western Joshua trees, though no quantitative data supports the observations. The analysis of the variables, however, shows the subjective observations to be accurate. Using data from nine study sites in the Southwest, eastern Joshua trees tend to be shorter with leaf rosettes concentrated below 3.5 m. This places a greater percentage of their biomass closer to the ground (Table 2). The differences between the two species are statistically significant for both variables ($P < 0.01$).

Joshua tree wildfire mortality on Cima Dome was greater than previously found in three *Y. brevifolia* populations in and near Joshua Tree National Park (Table 1). Eastern Joshua trees were more likely to be destroyed by wildfires than western Joshua trees, and the difference was significant ($P < 0.01$). For both species, root crown sprouting was overwhelmingly the most frequent survival mechanism though the frequency of this adaptation was greater for western Joshua trees.

It was of interest that there was a positive correlation ($r = 0.79$) between Joshua tree survival in the four studies and precipitation falling during the 12-month period after the fire. The correlation, however, was not statistically significant ($P = 0.21$).

### Discussion

The low survival rate of eastern Joshua trees after the Dome Fire, as depicted in and around the study site, diminishes the chance of a robust recovery when compared to sites where the western Joshua tree occurs. The relatively short, compact growth habit of *Y. jaegeriana* enables ground fires to more easily reach the crown and to concentrate damaging heat around the trunk base. This information may have unique management implications for agencies responsible for fire prevention and control in areas where eastern Joshua trees face significant wildfire threats.

### Table 1. Joshua tree post-fire survival rates. Top three studies analyzed western Joshua tree, *Yucca brevifolia*. Two bottom (current) studies examined eastern Joshua tree, *Y. jaegeriana*. Mean percentage survival by root crown sprouting for the two species were significantly different (independent two sample t-test, $P < 0.01$). Shaded area separates western Joshua tree data.

<table>
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<tr>
<th>Study</th>
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<th># Trees in study</th>
<th># Survived by sprouting (%)</th>
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<td>7</td>
<td>498</td>
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* Rounded to nearest whole percentage point

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Though beyond the purview of this study, the role of increased frequency and intensity of drought as well as an increase in annual temperature associated with climate change likely contributed to the high mortality found in this study. Unlike fires described in previous investigations, the Dome Fire was preceded by a decade where eight out of ten years experienced precipitation below the long-term average, the greatest frequency of drought in the California deserts since record keeping began (NOAA, 2021). In addition, mean annual temperature rose 2.2°C during this same period. Presumably, these factors resulted in a reduction in moisture content of trees making them more susceptible to fire damage (Vose et al., 2016). More importantly, the twelve months after the fire the burn area experienced more drought and another degree rise in temperature (NOAA, 2021). Though not statistically significant, it is of interest that a positive correlation \( r = 0.79 \) was found between the likelihood of survival and precipitation the year following a fire in the four studies examined here. Wildfires followed by drought may be more devastating to Joshua trees and additional investigations are needed to determine what role a changing climate may have in the severity of wildfires involving Joshua trees. Unfortunately, and with the data in hand, it is not possible to sort out the contributions made by the unique morphology of the eastern Joshua tree and the impact of factors associated with climate change.

This preliminary study did not examine the survival rates of shrub species known to be critically important nurse plants for seedling Joshua trees (Brittingham and Walker, 2000; Cornett, 1991, 2018). A cursory visual examination indicated few shrubs survived, a factor that would further exacerbate Joshua tree recovery prospects. The only shrub whose survival rate numbers were well over 50% was the trunkless banana yucca (Yucca baccata). This species has not, however, been recorded as a nurse plant for Yucca jaegeriana.

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**Literature cited**


Remote-sensed structural re-evaluation of Copper Mountain in the vicinity of Copper Mountain Community College, San Bernardino County, California

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¹California Professional Geologist and Certified Engineering Geologist
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Introduction
Copper Mountain Community College (CMCC) is a focal location for continuing and vocational education in the Yucca Valley, Joshua Tree and Twentynine Palms desert communities. It is also the meeting place for this year’s Desert Symposium. The college is nestled near the southwest flank of its namesake, Copper Mountain. The existing campus occupies approximately 45 acres in the northern portion of Assessor’s Parcel Number 0608-161-04. The larger parcel provides an opportunity for future growth of the college to eventually cover up to 117 acres (Figure 1).

Geomorphic Setting
Copper Mountain Community College is situated along the boundary of the Transverse Range and Mojave geomorphic provinces (Figure 2). The boundary between the two provinces is delineated by the west-striking Pinto Mountain Fault Zone. Although the Pinto Mountain Fault has not been the source of a historic earthquake, subsurface fault investigations have proven that the fault is active (Byerly, 2009). The fault is also considered to be active by the California Geological Survey (CGS) and the Southern California Earthquake Center (SCEC). CGS included the main strand of the fault within an Alquist-Priolo Earthquake Fault Zone (A-P EFZ), designated by the State to incorporate active and potentially active fault strands (California Division of Mines and Geology, 1988; Hart and Bryant, 1997; San Bernardino County, 1993). Copper Mountain Community College does not lie within the A-P EFZ associated with the Pinto Mountain Fault and no faults have been mapped on the campus or the remainder of the larger parcel. CGS and SCEC have determined that the Pinto Mountain Fault Zone may be capable of generating a Maximum Considered Earthquake (MCE) of magnitude M 7.5.

Tectonic Setting
The campus is separated from its namesake landform, Copper Mountain, by the Pinto Mountain Fault (Figure 3). The main strand of the fault is mapped approximately one-quarter mile north of the northern edge of the campus. Copper Mountain is a relatively small, east-west oriented hill located about one-half mile northeast of the college, near the Pinto Mountain Fault. Copper Mountain exhibits evidence of uplift as a pressure ridge and offset as a shutter ridge. In addition to Copper Mountain, the name...
is also applied to an adjacent, larger, northwest-trending ridge. For clarity, in this study, this larger, northwest-trending landform is informally referred to as “Copper Ridge,” to distinguish it from the smaller, west-oriented Copper Mountain.

“Copper Ridge” is bounded on the southwest by the northwest-striking Copper Mountain branch of the Calico-Hidalgo-Emerson Fault Zone and on the northeast by the northwest-striking Hidalgo branch of the fault zone (Figure 3). North-striking splays of the Emerson branch of the fault zone are located northwest of the campus. The Calico-Hidalgo-Emerson Fault Zone is part of the seismically active eastern California shear zone (ECSZ). The ECSZ has been the source of several large strike-slip earthquakes during the past thirty years, including the 1992 Landers, 1999 Hector Mine, and 2019 Ridgecrest-Trona earthquakes.

The section of the Copper Mountain Fault between Copper Mountain and “Copper Ridge” is not included within an A-P EFZ, unlike the fault farther northwest. The state has included the section of the Hidalgo fault bounding Hidalgo Mountain to the northwest within an A-P EFZ, but the portions of the fault northeast of “Copper Ridge” are not included within A-P Zones. The Emerson Fault northwest of Johnson Valley is included within an A-P EFZ, although sections of the fault northeast of the campus have not been so zoned.

This study concentrates on a re-evaluation of the tectonic and gravitational structure of Copper Mountain and “Copper Ridge.” Copper Mountain represents a symmetrical, doubly-plunging antiform. The antiformal shape suggests that the mountain may be underlain at relatively shallow depth by unmapped and unnamed west-striking reverse faults, with a south-dipping fault along the north flank and an opposing north-dipping fault along the south flank. The mountain has the geomorphic appearance of a “watermelon seed” squeezed up between the compressive “fingers” of these suspected reverse faults. Similarly, “Copper Ridge” geomorphically appears to have been uplifted along short, unmapped, northwest-striking reverse faults as well. In the vicinity of the college, the Pinto Mountain Fault offsets a shutter ridge of older alluvium immediately northeast of the campus.

**Geologic Setting**

Dibblee (1968a) and Jennings (1977) geologically mapped the rocks of Copper Mountain and “Copper Ridge” as biotite-rich and iron-stained quartz monzonite (Figures 4a and 4b). Dibblee mapped the surficial rocks in the western portion of Copper Mountain as hornblende diorite-gabbro. Although gneissic rocks are mapped southwest of the Copper Mountain fault, most of these older metamorphic rocks are found within “Copper Ridge.” Foliation mapped by Dibblee in the gneiss is generally oriented northeast-southwest, although foliations mapped north of Copper Mountain display a more steeply inclined arcuate pattern.

Rogers (1966), Dibblee (1968a) and Jennings (1975, 1992, 1994) identified several individual faults in the vicinity
of Copper Mountain, including the Pinto Mountain Fault. Although left unnamed by Dibblee, these faults included strands that would later be identified as the Copper Mountain and Hidalgo Faults. Dibblee did not recognize any landslides failing from either Copper Mountain or “Copper Ridge.” No landslides are mapped on the mountain or ridge in the San Bernardino County landslide database (San Bernardino County, undated). Sunfair Dry Lake (also referred to on some maps as Coyote Dry Lake) is mapped by Dibblee as Recent lacustrine sediments within a small depression northwest of Copper Mountain. A smaller, unnamed dry lake bed is also mapped by Dibblee as Recent lacustrine sediments within a fault step-over of the Hidalgo Fault, northeast of “Copper Ridge.” The presence of these lake beds adjacent to mountainous landforms suggests that these lake beds are owe their existence to transtensional and/or normal fault offset along mapped (and perhaps unmapped) strands of active faults bounding both of these lake beds.

Remote Sensing
A re-evaluation of the structure of Copper Mountain and the southeast portion of “Copper Ridge” was conducted utilizing virtual, time sequential, mosaicked, three-dimensional aerial imagery available from Google Earth (GE) on the Google Earth Pro platform. The dates of the aerial imagery ranged from 1985 to 2021. The evaluation utilized a digital geologic map KML overlay available from the U.S. Geological Survey (USGS) for Dibblee (1968a). Our evaluation also utilized USGS topographic maps, San Bernardino County Geologic Hazard Overlay maps, and the county’s digital Landslide Hazard Zones, draped over the Google Earth Pro DEM frame, to further identify and refine structural elements of Copper Mountain and “Copper Ridge.”

Our evaluation of draped maps utilized a vertically exaggerated (up to three times) digital elevation model (DEM) framework. Using the tilt and rotation tools available in Google Earth, the imagery was then reviewed at oblique angles, varying orientations of sun angle, and varying azimuth. This method of imagery analysis enhances tonal pixilation of the imagery from shadows, vegetation, and topography. Enhanced pixilation of the imagery, combined with enhanced topographic features, also allowed for interpretation of geomorphic and geologic structure within areas mapped as alluvium around the mountain and ridge. A three-fold vertically exaggerated image of the geologic map developed by Dibblee (1968a), tilted approximately 30˚ from the horizontal plane and rotated clockwise approximately 315˚ from due north, is provided as Figure 5 as an example.

Fault evaluation
An undated USGS fault database originally issued using ArcGIS, was converted to KML and adopted by San Bernardino County as the county’s Fault Hazard database, probably around 2004 (San Bernardino County, 2005).
The Pinto Mountain Fault has been re-examined along its length by the USGS (Menges, Matti, and Dudash, 2016; Langenheim and Powell, 2009; Hopson, 2012; and Matti, in press). The Pinto Mountain Fault displays left-lateral, strike-slip offset. The fault offsets a shutter ridge in Pleistocene-age sediments immediately north of CMCC. Although the Pinto Mountain Fault is mapped by Dibblee (1968a, 1968b) as a single trace in the vicinity of the campus, subsurface fault investigations between Yucca Valley and Twentynine Palms have documented a zone of older fault ruptures up to three-quarters of a mile in width (Byelry, 2009). A subsurface study conducted west of CMCC found more traces of the fault outside of the A-P Zone than in it (Byelry, 2009). The orientation of the Pinto Mountain Fault Zone is considered to control the orientation of Copper Mountain.

Lineament Analysis
Our evaluation of the Copper Mountain, Hidalgo, West Calico and Emerson Faults includes the mapped traces, and preliminary inclusion of geomorphically and tonally defined lineaments identified via remote sensing. More than 60 lineaments with topographic expression in alluvium were identified in the area around Copper Mountain and “Copper Ridge.” Locally, these topographic lineaments generally coincide with faults mapped by the USGS and CGS (Figure 6). These lineaments generally trend north and/or northwest, similar to the faults previously mapped in the Mojave Province. Our analysis found that these lineaments form an anastomosing pattern among the previously mapped faults. The anastomosing pattern is also evident on side-looking airborne radar imagery available from the USGS (USGS, 1985). If some or all of these lineaments represent active faulting, the pattern suggests that fault ruptures may exhibit complex fractal ruptures during large earthquakes similar to the complex pattern expressed during the Landers earthquake.

In addition to mapping digital and legacy fault locations available from the USGS, CGS, San Bernardino County, and Dibblee (1968a), our re-evaluation of the structure of Copper Mountain and “Copper Ridge” identified an additional approximately 600 lineaments. Since the analysis relied on mapping derived from remote sensed imagery, field truthing of the source of these structural elements was beyond the scope of this analysis. However, the topographic, geomorphic, and tonal character of these features is suggestive of faulting and/or landsliding. Our detailed analysis focused on the Copper Mountain and “Copper Ridge” landforms. Figure 7 presents the inclusive results of our lineament analysis.

Most of the lineaments observed in the immediate vicinity of Copper Mountain and “Copper Ridge” trend in northwesterly directions, parallel and sub-parallel to the Copper Mountain, Hidalgo, and West Calico fault zones. Figure 8 presents these northwest-oriented lineaments in two-dimensional plan view. Figure 9 displays the unretouched view of the imagery to the northwest at 3X vertical exaggeration, rotated approximately 315 degrees azimuth, and tilted obliquely at approximately 30 degrees from the horizontal plane, in order to accentuate the tonal, topographic, and geomorphic linear features. Figure 10 shows the same view with the lineaments mapped as blue lines following these linear trends, along with mapped faults.
A second prominent grouping of lineaments is oriented in a westerly direction, parallel to the Pinto Mountain Fault Zone. Several of these lineaments are identified in the vicinity of CMCC. A determination of the origin of these lineaments is beyond the scope of this investigation, but the proximity of vegetation and topographic lineaments oriented parallel to the Pinto Mountain Fault, combined with the width of the fault exposed in trenches east and west of the campus, suggests that several of the more prominent lineaments may represent potentially active fault traces. Since the campus lies entirely outside of the A-P EFZ associated with the fault, an A-P level investigation of any future placement of human occupancy structures on the parcel may not prove warranted. Still, due to the expected educational land use of the remainder of the parcel, further analysis of lineaments on or in the vicinity of the campus may be advisable should additional human occupancy structures be proposed on the site in the future. Figure 11 presents these west-oriented lineaments in plan view. Figure 12 displays the unretouched view of the imagery to the west at 3X vertical exaggeration, rotated ~315 degrees azimuth, and tilted obliquely at ~30 degrees from horizontal. Figure 13 shows the same view with the lineaments mapped as blue lines along these linear trends, along with mapped faults.

Review of the aerial imagery revealed a third set of lineaments trending in a northeasterly direction. These lineaments may represent faults antithetical to the predominant northwest-striking Copper Mountain and Calico-Hidalgo Fault Zones. The Kickapoo and Galway Lake Faults represented antithetical faults that ruptured during the Landers fault rupture sequence (Rockwell et al., 2000). Although no known active, northeast-striking faults have been mapped in the Copper Mountain area, the northeast-striking Morongo Valley Fault and southwestern-most section of the Pinto Mountain Fault are oriented northeasterly and demonstrate evidence for potentially active offset in Morongo Valley (Jordan, 2012). In addition, the active Manix Fault strikes to the east-northeast near Afton Canyon (Miller and Valin,

A fourth set of lineaments trends in a northerly direction. This set is oriented parallel to the local branches of the Hidalgo and Emerson faults east and west of “Copper Ridge,” respectively. These local branches are mapped by the USGS (2020) as north-striking. These lineaments are expressed geomorphically and topographically as boundaries between drainage patterns developed in the alluvial materials north of “Copper Ridge.” The lineaments are also parallel to the sections of the Landers Fault Zone and the Mesquite Lake Fault at the latitude of “Copper Ridge.”

Landslide evaluation

Dibblee (1968a) mapped the entirety of Copper Mountain and “Copper Ridge” as intact metamorphic and granitic bedrock (Figure 4a). The engineering community has long recognized that disturbed rock materials do not possess the strength characteristics of undisturbed bedrock and, instead, refer to geologic materials affected by landsliding as soils (Terzaghi, 1943; Terzaghi, Peck, and Mesri, 1996). Once a large landslide has failed from a bedrock slope, the reduction in strength of the disturbed materials provides discontinuities and additional shears for the introduction of rainfall into the rock and continued movement of the materials affected by landsliding (Varnes, 1978). This
A fractal characteristic of mass wasted material results in the subsequent failure of relatively younger and smaller, nested landslides from the initial landslide (Mandelbrot, 1983).

Our evaluation identified at least 150 large, deep-seated landslides failing from the slopes of Copper Mountain and “Copper Ridge” (Figure 20). In addition, numerous nested landslides were also identified and preliminarily mapped within the larger landslides. These smaller, nested landslides may have occurred during the late-Pleistocene to mid-Holocene. Due to the repetitious and fractal nature of these smaller landslides, only a few nested sets were mapped for this evaluation. An effort to map all of the smaller landslides recognized in the study area would have consumed time and resources far in excess of the scope of this study. Due to the fractal nature of landslides, the landslides mapped for this evaluation are considered to represent the minimum number of slope failures present on Copper Mountain and “Copper Ridge.”

Neither Dibblee (1968a) nor Jennings (1977) mapped landslides along the slopes of Copper Mountain or “Copper Ridge.” Our slope evaluation, coupled with an analysis of drainage patterns developed on, and below, the slopes, indicates that most, if not all, of Copper Mountain has been affected by deep-seated landsliding (Figure 20). Landslides on Copper Mountain generally fail perpendicular to the ridgeline of the mountain. Landslides also are recognized along the steep southwest flank of “Copper Ridge.” The removal of slope support due to strike-slip fault offset along the Copper Mountain Fault is considered the source of the steep gradient of the southwest flank. The northeast flank of the ridge is less steep. Landslides recognized along this flank are generally larger in area, but the largest landslides identified in this evaluation were recognized as facing toward the northwest from the high point of “Copper Ridge” (Figure 21).

Deep-seated landslides form large, coalescing landforms. The coalescing of similar-sized landslides suggests that these failure events occurred contemporaneously, if not simultaneously, and perhaps during severe seismic shaking associated with nearby faults, like the Pinto Mountain, Copper Mountain, Hidalgo and/or Emerson faults. These mapped potential seismic sources are shown on Figure 21. Alternatively, the coalesced masses may have formed during periods of greater precipitation, such as is expected to have occurred in the area during wet episodes of the late Pleistocene. Even more likely would be simultaneous strong ground shaking during wet periods.

Geomorphic evidence suggests that lateral fault offset along the Pinto Mountain Fault has resulted in the steep southern flank of Copper Mountain. Similarly, fault offset along the north flank of Copper Mountain appears to have resulted in the relatively steep northern flank of the mountain. The geomorphic evidence also suggests that lateral fault offset along the Copper Mountain Fault has resulted in the steep southwestern flank of “Copper Ridge.” All of these steep slopes are considered to represent fault scarps (Figures 22 and 23). The removal of slope support due to strike-slip fault offset resulted in the relatively rapid oversteepening of these slopes. Based on current fault mapping, the steepened northern flank of Copper Mountain would have been attributed to movement along the Copper Mountain Fault. However, geomorphology suggests that the northern flank should owe its steepened gradient to a previously unrecognized, parallel branch of the Pinto Mountain Fault along the
north flank of the mountain. Our lineament analysis identified at least two lineaments at this location that are suggestive of strike-slip fault offset.

The increase in the slopes of the south and north flanks of Copper Mountain and the southwest flank of “Copper Ridge” has resulted in mass wasting of these slopes (Figure 23). Increased surface runoff from the mountain and ridge flanks has increased the rate of erosion of these slopes. The increased rate of erosion further exacerbates more recent (Holocene) slope failures from these slopes. The evidence of recent landsliding along these slopes suggests that the most recent movement along both the Pinto Mountain and Copper Mountain Faults is also Holocene in age.

Conclusions

Previous geologic mapping of Copper Mountain, “Copper Ridge,” and the area around these landforms focused on the geologic rock units exposed in the uplifted rock masses. Geomorphic evidence for mass wasting events was not recognized during this previous mapping. Tectonic uplift of Copper Mountain and “Copper Ridge,” in combination with lateral offset associated with mapped strike-slip faults and deep-seated landslides, have produced the geomorphic landforms of the mountain and ridge. Geomorphic evidence for coalesced large deep-seated landslides and nested smaller, younger shallow-seated landslides are recognized on the Google Earth Pro imagery relied upon for this re-evaluation of structure. These were added to a structural geologic map of the area. The suspected Pleistocene age of these previously unrecognized larger landslides and the potential Holocene age of previously unrecognized younger landslides have implications for suspected faults traversing the mountain and nearby areas. Faults previously mapped traversing pre-Quaternary “bedrock” are in fact traversing Quaternary-age landslides that can be considered to be “sedimentary deposits.” The reclassification of “bedrock” as Quaternary-age deposits would change the states of activity of these faults from “not active” or “unknown age” to at least potentially active, and, perhaps, to active faulting.

Fault offsets on the bounding Pinto Mountain, Copper Mountain, and Calico-Hidalgo faults have exposed the flanks of the mountain and ridge to the concentrated effects of mass wasting and erosion. Geomorphic evidence for numerous vegetation and topographic lineaments traversing Copper Mountain, “Copper Ridge,” and the alluvium in the vicinity of the college, were identified on Google Earth Pro imagery and added to the structural geologic map of the area. Weak planar contacts between the overlying hornblende diorite and the underlying biotite quartz monzonite exposed in the mountain and ridge concentrate surface erosion along weakened rock pathways.

The re-evaluation of the geologic structure of the Copper Mountain area has significant implications for the anticipated future expansion of CMCC. Future geologic and geotechnical evaluations of potential hazards to facilities should include the potential for previously unmapped and unrecognized potentially active faults traversing the larger parcel, outside of the current boundaries of the A-P EFZ associated with the Pinto Mountain Fault. The anticipated presence of more faults may have implications for surface rupture events that could impact future educational structures.
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A Late Miocene camelid tracksite in the Muddy Creek Formation near Mesquite, Nevada

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ABSTRACT—We document a late Miocene (Hemphillian North American Land Mammal Age) camelid tracksite in the Muddy Creek Formation, near Flat Top Mesa, adjacent to Mesquite, Nevada. Through the use of photogrammetric modeling we are able to identify forty camelid tracks, Lamaichnum isp., impressed into a thin layer of calcareous quartz sandstone. The depositional environment is interpreted to have been an ephemeral lake within the fluvial-dominated Muddy Creek basin. Remnants of stalks and branches of rushes, cf. Juncus sp., are present. The average length of the camel tracks is 135 mm. They are larger on average than older tracks in the Mojave Desert region, which reinforces a previously recognized trend of increasing camelid body size through the Miocene.

Introduction

Camelid footprints are abundant in Neogene deposits of North America. They first appear in the Oligocene, and they range geographically from Mexico to Canada (Lucas and Hunt, 2007). Camel tracks are especially abundant in the Mojave Desert region where camelid morphological evolution is evident in fossil bones and trackways found in the Barstow, Avawatz, Horse Spring, and Muddy Creek formations, among others (Lockley and Hunt, 1995; Sarjeant and Reynolds, 1999; Jones, 2002; Lofgren et al., 2006; Lucas and Hunt, 2007). Lucas and Hunt (2007) identified a gradual increase through time in the average foot size in camel tracks preserved in these formations.

In this study, we document a previously undescribed fossil tracksite near Flat Top Mesa, adjacent to Mesquite, Nevada (Figs. 1, 2). This new tracksite contributes to the plethora of North American Neogene camelid footprint occurrences and helps build a more complete picture of camelid body size and behavior.

Camel evolution is well documented in the Cenozoic fossil record of North America, following trends similar to those seen in horses: Eocene ancestors with small body sizes and undifferentiated morphologies evolved into much larger animals with a reduced number of toes and other specialized features by the end of the Pliocene (Gauthier-Pilters and Dagg, 1981). Rapid radiation in the Miocene transformed camelids into the familiar taxa that we know today (Lucas and Hunt, 2007). The Miocene uptick in evolutionary pressure is generally interpreted to have been driven by the spread of grasslands (Janis, 1993). The camel foot evolved from having five toes to having two toes with nails and two pads connected by a septum (Arnautovic and Abdalla, 1969; Sarjeant and Reynolds, 1999) (Fig. 3). Camelid footprints are distinctive in that...
the anterior tips (or ‘toes’) are either parallel or divergent, whereas the digits of other artiodactyl toes are convergent (Sarjeant and Reynolds, 1999). The typical camelid footprint is heart shaped (Fig. 4). This results from the camel’s subdigitigrade mode of locomotion in which the animal walks on the palms of its feet, not just on its toes like other artiodactyls (Sarjeant and Reynolds, 1999) (Fig. 3).

A total of 40 camelid footprints occur within three exposures of the same bedding plane: Exposures A and B (Fig. 2), and Exposure C (Fig. 5). Exposure C occurs several meters downslope from Exposures A and B and contains only one truncated print. These exposures occur within a calcareous quartz sandstone bed that is mostly covered by a soil crust. The friability of this limestone makes removal of the tracks impossible. In this study we used a combination of 3D photogrammetric models and singular photographs to identify the ichnotaxon and to document and interpret the tracksite.

Age, geologic setting, and previous work
The Muddy Creek Formation was deposited during the Hemphillian North American Land Mammal Age (NALMA) (Reynolds and Lindsay, 1999), which extends from 4.8 Ma to 9.0 Ma (Lindsay et al., 2002; Tedford et al., 2004). In terms of the global time scale, this NALMA extends from late in the Miocene Epoch into the earliest Pliocene. Due to stratigraphic complexity and a dearth of datable horizons, the radiometric age of the Muddy Creek Formation is not tightly constrained (Muntean, 2012). The Flat Top Mesa tracksite occurs well below the top of the formation and is most probably late Miocene in age.

The Muddy Creek Formation is exposed throughout southeastern Nevada, from Moapa Valley to Mesquite. In some areas, the formation is unconformably capped by a prominent, mesa-forming, petrocalcic horizon that is typically 2–5 meters thick (Gardner, 1972; Williams, 1996; Brock and Buck, 2009). This petrocalcic interval forms Mormon Mesa, in eastern Clark County. Flat Top Mesa, near the tracksite, is an erosional outlier of the same resistant surface.

The depositional environment of the Muddy Creek Formation was fluvial and lacustrine, with small streams and lakes scattered across an expansive sandy plain (Williams, 1996). The strata in the Mesquite region are predominantly fluvial. These sediments originated in...
the Caliente Caldera complex, north of the field area, and they filled the Mesquite basin following Basin-and-Range extension (Kowallis and Everett, 1986; Forrester, 2009; Muntean, 2012). Deposition occurred in several small basins that eventually coalesced into a single large basin (Reynolds and Lindsay, 1999). Kowallis and Everett (1986) described the Muddy Creek Formation sediments in the Mesquite area as immature, basin-fill, mudstones, siltstones, and sandstones. They interpreted the environment to have included playas, with evaporites and long periods of standing water. Temperatures were slightly cooler with more rainfall than today, and the topography was less hilly.

Fossils within the Muddy Creek Formation include invertebrate trace fossils such as horizontal and vertical burrows, as well as mammal, bird, and lizard prints (Kowallis and Everett, 1986). Some vertebrate skeletal fossils have been reported from this formation, but none have been formally described (Williams, 1996; Howe, 1997; Reynolds and Lindsay, 1999). Other exposures of the Muddy Creek Formation contain avian, ursid, camelid, and felid tracks, typically occurring in sandstone (personal observation).

Methods

Photographs were taken following the protocol of Faulkingham et al. (2018). A Nikon D750 DSLR camera and Agisoft PhotoScan Professional software were used to take and process photogrammetric images. Separate photogrammetric models were made for exposures A and B. The single partial track in Exposure C was not examined photogrammetrically. Analysis of each footprint was based on the photogrammetric model, accompanied by individual photographs. Ichnotaxonomy follows Lucas and Hunt (2007). Thin sections of the trackway-bearing layer, oriented perpendicular to bedding, were prepared and examined microscopically.

Results and interpretations

When examining this tracksite visually, 19 prints are evident. However, analysis of the photogrammetric models revealed 21 additional prints, making a total of 40 prints (Figs. 8, 9). By turning the model over and analyzing the underside, print morphology is more easily seen. All footprints are concave epirelief (molds).

Track lengths range from 60 to 190 mm, with an average length of 135 mm (Fig. 10). The truncated print in Exposure C is not included in the size frequency histogram because not enough of the print is preserved to measure its length. The footprints are all subdigitrade and bidigital, with slight variations of a general heart shape. Posterior ends are round with a cleft, while anterior ends are pointed with parallel or diverging tips. The interdigital sulcus may or may not extend from the anterior cleft to the posterior cleft (Fig. 3). These differences are extramorphological variations which in many cases produced suboptimal prints. Such extramorphological variation may be caused by variable characteristics of the substrate or by different types of foot movement by the animals (e.g., walking versus running) (Peabody, 1948).

In modern artiodactyls, Murie (1974) demonstrated that extramorphological variation may produce different size and shape prints from the same foot.

Two examples of extramorphological variation in the Flat Top Mesa tracksite prints are illustrated in Figure 4. The presence of such variation at this site may have been caused by camels visiting the site at different times, when the substrate was wetter or drier. The single print in Exposure C (Fig. 5) reveals that when this deep track was made the substrate was soft and squishy enough for

Figure 6. Photomicrographs of a thin section of the track-bearing bed, a poorly sorted quartz sandstone in a carbonate matrix. The thin section was oriented perpendicular to bedding. A. Plane light. B. Crossed polars.

Figure 7. A. Cylindrical stalk of a macrophyte, cf. Juncus sp., encrusted with carbonate (in Exposure A). B. Small branched plant fragments encrusted with carbonate, Exposure A, adjacent to the cylindrical stalk. C. Photograph of Juncus effusus, for comparison with A and B ['Juncus effusus (common rush).’ Everwilde.com/store/juncus-effusus-Seed.html. January 15, 2022.]
the track to depress the light-colored trackway layer into the underlying, darker sediment. In contrast, many of the tracks in exposures A and B are so subtle that they were detected only through photogrammetry.

Depositional environment

The track-bearing horizon is a yellowish gray (Geol. Soc. Amer. color chart 5Y 8/1), calcareous, quartz sandstone, 2 to 4 cm thick (Fig. 5). In thin section, this layer is seen to consist predominantly of subrounded-to-angular, quartz grains, with rare grains of feldspar (Fig. 6). The grains are poorly sorted, ranging in size from 0.1 to 1.0 mm (fine to coarse sand), in a carbonate matrix. Some of the grains appear to be “floating” in the carbonate matrix (Fig. 6B).

Plant fossils are present in the form of cylindrical structures ~1.5 cm in diameter (Fig. 7A) as well as smaller, diverging branches (Fig. 7B). In size and morphology, these structures closely match those of the stalk and branching inflorescence of the rush Juncus sp. (Fig., 7C), which is a very common plant in modern freshwater environments. In North America, Juncus occurs as early as the Eocene, in the Florissant Fossil Flora of Colorado (Panjabi and Anderson, 2002).

We interpret the depositional environment of the Flat Top Mesa tracksite to have been a shallow pond with abundant emergent vegetation, including rushes. The camels presumably came to the pond to drink, but they would not have been able to utilize the aquatic plants as a food source; much of the carbon within the tissues of such aquatic macrophytes is not readily digestible by mammals (Mann, 1988).

The carbonate matrix of the trackway layer we interpret to be the product of photosynthesis-induced precipitation of carbonate in a hard-water lake. In such a body of water, photosynthesis by algae and rooted aquatic vegetation removes CO$_2$ from the water column, triggering the inorganic precipitation of CaCO$_3$ (Dean and Fouch, 1983). The stalks and branches of macrophytes thus became encrusted with carbonate (Fig. 7). The poorly sorted sand grains within the track-bearing layer were presumably introduced into the lake by fluvial processes. We suggest that a low-gradient stream was depositing sand grains into the lake at a slow rate, allowing CaCO$_3$ to precipitate contemporaneously with the deposition of the grains. Macrophytic vegetation, which may have been dense along much of the shoreline, would have been less dense where active sedimentation was occurring, at the mouth of a stream.

Laporte and Behrensmeyer (1980) observed that the set of conditions that must be met for vertebrate tracks to be preserved in the rock record includes relatively vegetation-free, moist sediment that becomes covered by a protective layer soon after the tracks are made. These conditions, they concluded, are most frequently met near the edge of a lake, or where a stream enters a lake. The sedimentology of the Flat Top Mesa tracksite matches this environmental description.
Systematic ichnology

Ichnogenus *Lamaichnum* Aramayo and Bianco, 1987

Material referred: 40 Footprints from the Flat Top Mesa tracksite near Mesquite, NV.

Lucas and Hunt (2007) simplified camel ichnotaxonomy into one ichnogenus and two ichnospecies (*L. guanico* and *L. macropodum*). These two ichnospecies are different only in footprint length, with *L. guanico* being shorter than 160 mm and *L. macropodum* ranging in length from 160 to 260 mm. Because size is not a reliable way to differentiate between ichnospecies, Lucas and Hunt (2007) acknowledged that camelid ichnotaxonomy is a work in progress. The Flat Top Mesa tracksite footprints range in length from 60 mm to 190 mm, with an average length of 135 mm (Fig. 8). Differentiating these ichnospecies provides no meaningful distinction in the case of this assemblage of tracks, so we refer them to *Lamaichnum* isp.

Discussion

By the mid-Miocene, North American camelid foot morphology had evolved into the shape that we see in modern camels (Lucas and Hunt, 2007). Following this trend, one would expect to see a slight increase in average footprint size from middle Miocene (Barstow and Avawatz formations) to late Miocene–early Pliocene (Muddy Creek Formation), by which time camels had evolved into the larger body type that is familiar in extant dromedary and bactrian camels. The trend toward generally larger footprints in the Mojave region can be seen in Figure 10, with the Muddy Creek Formation having a broader range in size distribution of footprint length than the Barstow and Avawatz formations. These size distributions in the older Barstow and Avawatz formations approximate normal (bell-shaped) distributions with an average footprint length of 90 mm (Fig. 8). The size distribution in the Muddy Creek Formation, although bimodal, also approximates a normal distribution but with a much wider range of sizes and a larger average track length of 135 mm. This bi-modal aspect could be a result of both juvenile and adult trackmakers being present. The Muddy Creek sample size (n = 39) is much smaller than that of the Barstow (n = 297). The number of measured specimens in the Avawatz was not specifically stated by Lucas and Hunt (2007); the estimation of n = ~ 48 is derived from Lucas and Hunt (2007, fig. 11). More Muddy Creek footprints need to be measured for a more robust comparison.

While the trend of larger footprints through time is evident within the Miocene of southwestern North America, it does not continue uniformly into the Pliocene and Pleistocene. In Anza-Borrego Desert State Park, for example, post-Miocene camelid lineages include some very large taxa, such as *Gigantocamelus* (Webb et al., 2006), which was the likely trackmaker of tracks as long as 270 mm (Remeika, 2006), but small camelid taxa were also present at Anza Borrego. Body fossils of at least five species of Blancan (Pliocene) and Irvingtonian (early Pleistocene) camelids have been recovered from the fossil beds of Anza Borrego (Webb et al., 2006); two are small and three are large (G. Jefferson, pers. commun. 2022).

Modern camels have a distinctive walking gait in which they walk with both right legs in succession, followed by both left legs. This is known as a pacing gait (Webb, 1972) and also as a lateral sequence gait (Hildebrand, 1980). This gait helps conserve energy by permitting longer
strides and fewer steps, at the expense of not being able to make quick directional changes (Sarjeant and Reynolds, 1999). The pacing gait produces a trackway with the slightly larger manus in front of the slightly smaller pes (Webb, 1972). This is demonstrated in a famous Barstow Formation trackway made by two camelsid pacing side by side, on display at the Raymond Alf Museum (Sarjeant and Reynolds, 1999, fig. 9). At the Flat Top Mesa tracksite, there are two possible examples of a pacing gait. One possible example in Exposure A is recorded in prints A1/ A5/A8 (Fig. 8); another possible example is recorded in prints B1/B2/B3/B4 in Exposure B (Fig. 9).

Possible trackmakers for the Flat Top Mesa tracksite include Megatylopus and Alforjas. Skeletal fossils of these genera have been reported from the upper Muddy Creek Formation near Mesquite (Howe, 1997; Reynolds and Lindsey, 1999), but they have not been formally described. Megatylopus is also found in the Panaca Formation, a Hemphillian NALMA deposit with a large fossil assemblage (Reynolds and Lindsey, 1999; Meyers, 2011).

Ethologically, not much can be said about the behavior of the camels that left these impressions. However, we can make a few assumptions given the substrate and the orientation of the prints. While exposures A and B contain tracks headed in different directions, in general, these camels were traveling northward. In Exposure A, most of the prints are heading northeastward (Fig. 8), while in Exposure B, the prints are generally heading northwestward (Fig. 9). Determination of adults vs. juveniles was not attempted due to the small sample size, as well as the extramorphological variation of the prints. The presence of multiple individuals of various ages is probable because prints occur in a range of sizes. There is only one instance of overprinting (one print on top of another), with A11 and A12 (Fig. 8), which suggests a relatively low level of activity within the area at the time the sandstone was deposited. The calcareous sandstone represents a relatively long-standing body of water, possibly making it a reliable water source for the camels to return to.

It is not possible to determine whether all of these prints were imprinted at the same time. It is possible that different groups passed through the site, days or even hours apart (cf. Lockley and Hunt, 1995). The tracks could represent a group of individuals, or a pair of individuals passing through at different times, or even several individuals passing through one at a time, separated by hours or days. However, the occurrence of tracks made by multiple individuals headed in the same general direction suggests that these animals were gregarious, and they were members of a small herd. Wild dromedary and bactrian camels travel in small herds, especially the females with their young (Chen and Yuan, 1979; Spencer et al., 2014), so it is reasonable to assume that these Miocene camels did also. Assuming these prints record a herd of camels visiting the lake at the same time, the maximum number of individuals at the Flat Top Mesa tracksite would be around 20. Exposure A has three possible trackways (A1, A5, A8; A2, A4; and A22, A14, A15). Exposure B has one possible trackway (B1-B4). All the rest could be separate individuals.

Conclusions
The Flat Top Mesa tracksite contributes new data to the Neogene camelid footprint record in southwestern U.S.A. The site consists of 40 camelid footprint molds (concave epirelief) identified as Lamaichnum isp., with an average length of 135 mm. Footprint size follows the general camel evolutionary pattern; on average the tracks are larger than earlier Miocene footprints.

It is probable that the Flat Top Mesa tracksite records the presence of multiple individuals, probably in a small herd, and possibly with juveniles present. We suggest that these animals came to a shallow pond at a point where a low-gradient stream flowed into the pond. They drank but did not linger. The Muddy Creek Formation is emerging as an important mammalian ichnological resource similar to other Mojave Desert formations.

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References


Review of Duchesnean and earlier Chadronian North American Land Mammal Ages and stratigraphic occurrences of Protitanops (Mammalia, Perissodactyla, Brontotheriidae) from late Duchesnean strata in Death Valley National Park, California, Trans-Pecos Texas, USA, and northeastern Chihuahua, Mexico

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ABSTRACT—The monotypic brontotheriid Protitanops is restricted to assemblages of late Duchesnean (= Du2) age. Protitanops curryi occurs in (1) the late early Du2 Titus Canyon Fauna from the Lower Red Beds of the Titus Canyon Formation (Fm) in Death Valley National Park, southeastern California, (2) the middle early Du2 Lower Porvenir Local Fauna (LF), the late early Du2 Upper Porvenir LF, and the late Du2 Upper Little Egypt LF of the Chambers Tuff Fm in Trans-Pecos Texas, (3) unnamed middle early to late Du2 assemblages from the lower part of the Chisos Fm in Big Bend National Park, Trans-Pecos Texas, USA, and (4) the middle early or late early Du2 Upper Rancho Gaitan LF from the Upper Conglomerate Member (Mbr) of the Prietos Fm in northeastern Chihuahua, Mexico. The Protitanops biochron is constrained by single-crystal laser-fusion (SCLF) 40Ar/39Ar dates of 37.68 and 37.15 Ma for units that bracket the Chambers Tuff Fm. Based on the Geologic Time Scale 2020, the dates indicate that the biochron is earliest late Eocene in age. Protitanops curryi is designated as an index taxon for middle early to late Du2. However, it is not recorded from the type or earliest Du2 Halfway and Lapoint Faunas from the Duchesne River Fm’s Dry Gulch Creek and overlying Lapoint Mbrs, respectively, the upper two of three units constituting the Duchesnean North American Land Mammal Age stratotype in northeastern Utah. The Lapoint Fauna underlies or brackets a tuff bed that is 39.47 Ma old, based on SCLF dating analysis.

Introduction

The monotypic brontotheriine brontotheriid Protitanops is found in fossil land mammal assemblages of late Duchesnean (Du2) age from the Titus Canyon Formation (Fm) of southeastern California, the Chambers Tuff Fm and the lower part of the Chisos Fm in Trans-Pecos Texas, and the Prietos Fm of northeastern Chihuahua, Mexico (Fig. 1). Protitanops curryi is characterized, in part, by (1) its comparatively large body size, bulbous paired frontonasal horns with oval cross sections, deeply saddle-shaped cranium, and swollen zygomatic arches, and (2) its retention of upper and lower post-canine diastemata and three lower incisor pairs (Fig. 2, Stock 1936a:plates 1–2, Wilson 1977:fig. 5, Mihlbachler 2008:298, figs. 147–148, Mihlbachler and Prothero 2021:7, figs. 3–10). Among other things, this paper reviews the biostratigraphy, biochronology, and geochronology of P. curryi. It also reviews the biochronologies and boundaries of subages constituting the Duchesnean and earlier Chadronian North American Land Mammal Ages. Doing so provides a framework for constraining the biochron of each P. curryi occurrence.
Abbreviations

Ar/Ar = 40argon/39argon, Ch1 = earliest Chadronian of Prothero and Emry (1996:676) = late Du2 or latest Duchesnean herein, Ch2 = late early, (earlier part of) middle, and (earlier part of) late Chadronian of Prothero and Emry (1996:676–677) = early Chadronian herein, Ch3 = (later part of) middle Chadronian of Prothero and Emry (1996:676–677) = middle Chadronian herein, Ch4 = (later part of) late Chadronian of Prothero and Emry (1996:676–677) = late Chadronian herein, CIT = California Institute of Technology (collection now at LACM), CM = Carnegie Museum of Natural History Section of Vertebrate Paleontology, Du1 = early Duchesnean, Du2 = late Duchesnean (includes Ch1 = earliest Chadronian = late Du2 herein), F:AM = Frick: American Mammals (collection now at American Museum of Natural History), FCs = Fish Canyon Tuff sanidine, FM = Formation, FMNH = Field Museum of Natural History Fossil Mammal Collection, ft = feet, HLSO = highest local stratigraphic occurrence, K-Ar = 40potassium/39argon, km = kilometer, LACM = Natural History Museum of Los Angeles County Vertebrate Paleontology Department, LF = local fauna, LLSO = lowest local stratigraphic occurrence, loc = locality, m = meter, Ma = megannums, Mbr = Member, mm = millimeter, MMhb-1 = McClure Mountain hornblende, n. sp. = new species, NALMA = North American Land Mammal Age, Or1 = earliest Orellan, Or2 = late early Orellan, RFAD = regional first appearance datum, RLAD = regional last appearance datum, SCLF = single-crystal laser-fusion, SDSNH = San Diego Society of Natural History (San Diego Museum of Natural History Department of Paleontology), TMM = Texas Memorial Museum Vertebrate Paleontology

Methods

The Uintan, Duchesnean, Chadronian, and Orellan NALMAs are after Wood et al. (1941:10–11, plate 1). Subdivisions of Du2 (includes Ch1 herein as late Du2; see below) follow Lander (2019a:150), but now include some modifications based on updated taxonomic, biostratigraphic, biochronologic, and geochronologic data presented below and in Figure 3. Uintan to Orellan Subage subdivisions and abbreviations (e.g., Du1 for early Duchesnean NALMA) follow Prothero and Emry (1996:674–678) as modified by Speijer et al. (2020:fig. 28.12, p. 1114) and Lander (2022:fig. 7, table 1) and substantially so in this report (see below). Lithostratigraphic units of the Titus Canyon, Chambers Tuff, Prietos, Duchesne River, and Chadron Fms are from Stock and Bode (1935:plate 2) and Lander (2019a:fig. 3), Wilson (1978:figs. 5, 7–9), Ferrusquía-Villafranca (1969:fig. 2, "unnamed lower formation") and Heiken (1971:fig. 3), Andersen and Picard (1972:fig. 1), and Clark (1967:fig. 2), respectively. According to Andersen and Picard (1972, 4, 10, 12–15, fig. 1), their four mbrs of the Duchesne River Fm (and equivalent horizons or mbrs of Kay 1934:359, plate 46, 1953:24, 1957:114) include (1) the Brennan Basin Mbr (= Randlett Horizon and lower third of Halfway Horizon), (2) the Dry Gulch Creek Mbr (= upper two thirds of Halfway Horizon), (3) the Lapoint Mbr (= lower
part of Lapoint Horizon, and (4) the Starr Flat Mbr (= uppermost part of Lapoint Horizon and overlying, previously unassigned strata). The horizons produced the respective Randlett, Halfway, and Lapoint Faunas.

As necessary, Ar/Ar dates were corrected using ArArReCalc (McLean 2009) to account for the currently accepted, astronomically calibrated Ar/Ar date of 28.201 Ma reported by Kuiper et al. (2008:501) for the FCTs flux monitor. All errors are reported at one sigma. Field efforts were conducted under National Park Service Scientific Research and Collecting Permit No. DEVA-2017-SCI-0037, which was issued to Nyborg (Principal Investigator), Lander (Co-Investigator), and Dr. Kevin E, Nick (Loma Linda University, Co-Investigator).

Biochronologic background

Uintan-Duchesnean (= Ui3-Du1) NALMA boundary. The Brennan Basin Mbr represents the lowermost unit of the Duchesne River Fm and the Duchesnean NALMA stratotype in the Uinta Basin of northeastern Utah. Contrary to Lander (2013:47, fig. C-1), the type specimen for the agriochoerid Diplobunops crassus (CM 2967) was probably found at a level above the brown sandstone bed that (Peterson and Kay 1931:295) constitutes the basal 3.0–6.1 m (10–20 ft) of the Randlett Horizon or Brennan Basin Mbr and the Duchesne River Fm at Randlett Point, between the Baeser and Leota Bends of the Green River. CM 2967 was found only 1.6 km (1 mi) south of Baeser Bend (Scott 1945:235) and, therefore, upsection and at least 8.0 km (5 mi) north of the contact between the Duchesne River Fm and underlying Uinta C (= Myton Mbr of Uinta Fm), as mapped by Kay (1934:plate 45) and Sprinkel (2007:plate 1) (see U.S. Geological Survey Vernal, Utah, Quadrangle, 1917 edition, scale 1:125,000). Consequently, Diplobunops crassus is probably a member of the Du1 Upper Randlett Fauna instead of the late Ui3 Lower Randlett Fauna (Fig. 3), which shares very nearly all of its species with the type early Ui3 Myton Fauna from Uinta C. Smaller-bodied Diplobunops matthewi is a member of the Myton Fauna from Myton Pocket, Leota Ranch, and other localities in the Uinta Basin. The P2–M3 length of the D. crassus type specimen is 13.3% greater than the mean P2–M3 length for D. matthewi from Uinta C (see Lander and Hanson 2006:fig. 2, table 1). Because it is easily distinguishable from latest Uintan D. matthewi

Figure 3. Biochrons and occurrences of age-diagnostic land mammal species in assemblages of Du1 to earliest Ch2 age, some of which contain Protitanops curryi.
on the basis of P2–M3 lengths alone and is represented, sometimes rather commonly, in assemblages previously reported to be of Du1 age, the RFAD of *D. crassus* in the Brennan Basin Mbr is defined herein as the Uintan-Duchesnean NALMA boundary. Other Du1 assemblages containing *D. crassus* with comparable P2–M3 or mean P2–M3 lengths include (1) the Skyline LF from the Skyline Channels at the base of Bandera Mesa Mbr of the Devil’s Graveyard Fm in Trans-Pecos Texas and (2) the Hancock Mammal Quarry LF from Unit E of the Clarino Fm in the John Day Basin of north-central Oregon (Lander and Hanson 2006:24–25, fig. 2, table 1). On the other hand, *Diplobunops crassus* also represents a substantially smaller-bodied species than *D. n. sp. (very large) of earliest and middle early Du2 age (see below). The M3 of *D. n. sp. (very large) (USNM 299571) is more primitive than in early late Ch2 (not Ch3; see below) *D. maximus* of the Little Pipestone Creek LF from the Climbing Arrow Mbr of the Renova Fm in southwestern Wyoming (see Douglass 1902:plate 9, fig. 4, Emry 1981:fig.1).

**Du1-Du2 boundary.** Lucas (1992:93) merged the type Duchesnean Halfway and Lapoint Faunas from the Dry Gulch Creek and overlying Lapoint Mbrs, respectively. The mbrs represent the middle two of four units constituting the Duchesne River Fm and the upper two of three making up the Duchesne NALMA stratotype in the Uinta Basin of northeastern Utah. The Lapoint Fauna was the basis for the La Pointian or Lapointean Subage of Clark and Beerbower (1967:59, fig. 24) and Lucas (2015:154, fig. 7), respectively. In contrast to earlier workers, Kelly et al. (2012:table 9) regarded the Halfway/Lapoint Fauna informally as middle Duchesnean in age. However, many species they recorded from the Lapoint Fauna also occur in the early Du2 or Viejan Porvenir LF, which, according to Wilson et al. (1968:595), is found in the lower third of the Chambers Tuff Fm in Trans-Pecos Texas. The Viejan Subage of Clark and Beerbower (1967:59, fig. 24) and Lucas (2015:154, fig. 7) was based on the Vieja Fauna (Porvenir LF; Wilson 1978:35). The Lapoint Fauna and Porvenir LF currently share at least seven species, as recognized herein. They include (1) the hyaenodontid *Neohyaenodon n. sp. (medium)*, (2) the brontotheriid *Duchesneodus uintensis* (monotypic genus), (3) the helatedid *Colodon stovalli*, (4) the hyracodontid *Hyracodon medius* (= *Mesamynodon medius* Peterson 1932:71 = *Amyodon medius* of Kay 1953:24 = *H. primus* Peterson 1934:388; Scott 1945:248, Rasmussen et al. 1999:423–424, this report), (5) the rhinocerotid *Penetrigonias hudsoni*, (6) the entelodontid *Brachyhyops wyomingensis*, and (7) the protoceratid *Poabromylus kayi* (Fig. 3, Wilson 1984:203, Lander 2019a:table 3). Six of the species in the Lapoint Fauna represent RFADs (bolded names), but four of those are also generic RFADs (underlined names) (Fig. 3). Six of the species exclusive of *Duchesneodus uintensis* are shared with the Lower Porvenir LF from the Big Red Horizon (i.e., below Lower Marker Bed), the lowermost unit of the Chambers Tuff Fm (Fig. 3, Lander 2019a:table 3; see Wilson 1978:figs. 5, 7, tables 4–5, 1984:201). In addition, *Hyracodon medius* plus the agriochoerid *Diplobunops n. sp. (very large)* (= *D. “maximus” = "Agriochoerus maximus" of Lander 2012) are found in an early Du2 correlative of the Lower Porvenir LF, the Lower Rancho Gaitan LF from the Upper Tuffaceous Mbr of the Prietos Fm at loc 2 of Ferrusquia-Villafranca (1969:plate 1) in the Cerros Prietos area ca. 32 km (20 mi) northwest of Ojinaga in northeastern Chihuahua, Mexico (Fig. 3, Lander 2019b:209). On the other hand, five of the taxa listed as shared with the Lower Porvenir LF (Neohyaenodon n. sp. [medium], *Colodon stovalli* [RLAD], *Hyracodon medius* [RLAD], *Penetrigonias hudsoni*, *Poabromylus kayi*) plus the generic RLAD of *Duchesneodus uintensis* also occur in the late early Du2 Upper Porvenir LF from the Blue Cliff Horizon (i.e., above Lower Marker Bed in lower [but not lowermost] part of Chambers Tuff Fm) (Fig. 3, Lander 2019a:147, table 3; see Wilson 1978:table 6, Mihlbachler 2008:316). Because it shares at least eight species with three assemblages of early Du2 age, the Lapoint Fauna is considered to be that age, as well, rather than middle Duchesnean (Fig. 3). Similarly and for reasons discussed below, Lander (2019a:150) regarded the Little Egypt LF from the uppermost part of the Chambers Tuff Fm as late Du2 rather than Ch1 in age and, accordingly, the Upper Porvenir LF (and, therefore, late Viejan Subage) as late early Du2. Correspondingly, the Lower Porvenir LF and the early Viejan Subage are viewed herein as middle early instead of earliest Du2 in age, whereas the Lapoint Fauna and, therefore, the Lapointean Subage are assigned to earliest Du2. In addition to RFADs cited above, the generic RLADs of the equid *Ephippus intermedius* (= *Duchesnehippus intermedius*)(uncataloged CM specimen, partial lower cheek tooth; see Kay 1953:24, 1957:114, Rasmussen et al. 1999:254, table 1) and the amynodontid *Megalamynodon sp. (= M. regulis* of this report, FMNH PM 22412) also occur in the Lapoint Fauna (Fig. 3).

A specimen of *Poabromylus kayi* (CM 11856) was found in the Lapoint Horizon 3.2 km (2 mi) southwest of Little Mountain and, therefore, probably represents the HLSO of an identifiable land specimen from the Duchesne River Fm. If geographic locality data are accurate, CM 11856 is probably from a level high in the Lapoint Mbr or perhaps even a level low in the overlying Starr Flat Mbr (see Sprinkel 2007:plate 1, Webb 2017:plate 1, respectively). If from the latter unit, it would represent the only identifiable land mammal specimen from the Starr Flat Mbr and a record of *Poabromylus kayi* possibly even as young as the one (FMNH PM 455) in the late early Du2 Upper Porvenir LF from the Blue Cliff Horizon of the Chambers Tuff Fm.

The Halfway Fauna, which occurs in the Dry Gulch Creek Mbr, also shares six species with the Lapoint Fauna. Records from the Halfway Fauna include (1) the cylindrodonid *Pareumys guensburgi* from SDSNH loc 3939, (2) the eomyid *Protadjidaumo typus* from SDSNH
locus 5939 and 6087, (3) *Epihippus intermedius* (CM 11845, type specimen, mandibular symphysis and right ramus with i1–m2), (4) the amynodontid *Megalamynodon regalis* (type [CM 11953] and topotype specimens [CM 9961, 11958]), (5) *Diplobunops* n. sp. (very large) (= D. "maximus" = "Agrichoerus maximus"); and (6) *Poabromylus kayi* from SDSNH loc 6339 (not 6389) (Fig. 3; see Kelly et al. 2012:table 6). The *Megalamynodon regalis* specimens probably represent the "amynodont" record shown near the middle of the Halfway Mbr by Kay (1934:plate 46; see 1953:24, 1957:114). However, species were subsequently assigned incorrectly to the Randlett Mbr by Scott (1945:251). USNM 299571, a specimen of *Diplobunops* n. sp. (very large) originally assigned to the lower part of the Lapoint Mbr by Emry (1981:563), was recently reassigned to the uppermost part of the Dry Gulch Creek Mbr at Halfway Hollow Quarry by Jensen et al. (2020:21, fig. 2b). Consequently, CM 26051 from the Red Narrows represents the only record of the taxon from the Lapoint Mbr (see Lander and Hanson 2006:28). The three SDSNH localities lie at about the middle of the Dry Gulch Creek Mbr (Kelly et al. 2012:209, fig. 14).

The listed species from the Halfway Fauna all represent LLSOs and, except for *Pareumys guensburgi*, RFADs, as well. In contrast, (1) the heseliohaco *Passaliscomys* sp., (2) the eomyid *Adjidaumo cf. craigi*, and (3) the eutypomyid *Eutypomys*, all from SDSNH locs 5939 and 6087 (Kelly et al. 2012:213, 220, 229–231, 241, figs. 12, 14, table 6), along with (4) *Megalamynodon regalis*, also correspond to generic RFADs. The indeterminate brontotheriid specimen from the Halfway Mbr that was assigned questionably to *Duchesnusoidius intermedius* by Kelly et al. (2012:230 table 6) (CM 11996) was also identified as the brontotheriid *Prototitanotherium* sp. by Scott (1945:245). No definitive occurrence of either taxon is currently recorded from the Dry Gulch Creek Mbr or the Halfway Fauna. Only *Diplobunops* n. sp. (very large) and *Poabromylus kayi* of the Halfway Fauna are shared with the middle early Du2 Lower Rancho Gaitan LF on one hand and the middle early and late early Du2 Lower and Upper Porvenir LFs on the other (Fig. 3, Wilson 1978:tables 4, 13, but *P. kayi* [FMNH PM 455] not included with Upper Porvenir LF in his table 6, 1984:201). However, the Halfway Fauna, like the overlying Lapoint Fauna, is also regarded in this report as earliest Du2 in age because they have at least six species in common.

The RFAD of *Epihippus intermedius* (= generic RFAD for *Duchesnushippus intermedius*) is from Halfway Hollow, 1.6 km (1 mi) south of Vernal-Lapoint Road (= Utah State Route 121) (Peterson 1932:67). That record appears to represent the oldest record of the Halfway Fauna from the Dry Gulch Creek Mbr (see Sprinkel 2007:plate 1, Webb 2017:plate 1). The respective level is perhaps below those of SDSNH locs 5939, 6087, and 6339, which (Kelly et al. 2012:209) lie near the middle of the mbr. Therefore, the RFAD of *E. intermedius* in the Dry Gulch Creek Mbr is tentatively used herein to define the Du1-Du2 boundary.

**Earliest-middle early Du2 boundary.** Wilson (1978:tables 4–5, 7, figs. 7–8) recorded a number of age-diagnostic land mammal species in the middle early Du2 Lower Porvenir LF from the lowermost part of the Chambers Tuff Fm in Trans-Pecos Texas that represented RFADs. The fossil-bearing interval, which is bracketed by the Buckshot Ignimbrite below and the Lower Marker Bed above, is identified as the Big Red Horizon in the Sierra Vieja, but not at Adobe Spring to the southeast (Wilson 1978:figs. 7–8). Among the species with RFADs in the Lower Porvenir LF are the ischyromyid *Quadratomus? gigans* and the rhinocerotid *Teletaceras mortvallisi* (Fig. 3). However, a number of species also represent generic RFADs. Among those taxa are the amphiid *Daphoenictis* n. sp. (small), the equid *Mesohippus viejensis* (includes *M. texanus*; Mason 1988:209, Lander 2019a:146), the agiochoroid *Agrichoerus* n. sp. A (medium), and the leptomerycid *Hidrotherium transpecosensis* (includes *Leptomeryx defordi*; see Prothero 1996:612, but *H. transpecosensis* has page priority), as indicated in Figure 3. Species with RLADs in the Lower Porvenir LF include the entelodontid *Brachyhippus wyomingensis*, whereas the oreodontids *Acistomyceter* (i.e., *A. middletoni*) and *Prodesmatochoerus* n. sp. A (large) are restricted to middle early Du2 (Fig. 3).

*Daphoenictis* n. sp. (small) is the same as canid sp. of Stock (1949:233), *Daphoeninae* genus and species indeterminate of Gustafson (1986:51), and *D. n. sp.* of Lander (2019a:146). It’s mean m1 and m2 lengths (13.0 and 7.0 mm, respectively) are 16.7% and 13.5% less than in *D. tedfordi* and it retains the m2 metaconid (Stock 1949:233, see p. 234, Gustafson 1986:53, table 13, TMM 40203-16, 40688-72). *Daphoenictis tedfordi* is considered herein to be of early late Ch2–Ch4 (not Ch3–Ch4) age (see below).

*Agrichoerus* n. sp. A (medium) is equivalent to *A. n. sp.* of Lander (1998:409, loc SB44b), which, in turn, includes *A. antiquus* of Wilson (1971a:31). It’s mean P2–M 3 length is 14.8% less than the P2–M3 length for an individual of early late Ch2 A. n. sp. B (large) from 15.2–15.8 m (50–52 ft) below Ash B of the White River Fm at Flagstaff Rim in central Wyoming (see Lander and Hanson 2006:fig. 2, table 1, p. 23). When compared to latest Ch2 A. n. sp. C (medium), (1) the P1 of A. n. sp. A (medium) is larger and seemingly always present, (2) the P4 is less molariform (e.g., P4 hypocone still incompletely selenodont), (3) the M1 sometimes retains the paraconule, and (4) the tympanic bulla is substantially less inflated (E. B. Lander unpub. data, Wilson 1971a:31, table 18, see figs. 11a, 18). *Agrichoerus* n. sp. B (large) is regarded in this report as early late Ch2 instead of Ch2–Ch3 in age, whereas A. n. sp. C (medium) is considered to be of latest Ch2–C3 age (see below). The latter species is the first member of the genus in which upper cheek tooth and basifrenal anatomies have attained structural grades typified by Or1 *A. antiquus* and the longer-ranging *A. antiquus* species group.
Prodesmatochoerus n. sp. A (large) is the same as P. n. sp. A of Lander (2019a:150) and Merycoidodontidae genus and species indeterminate no. 1 of Wilson (1971a:46). The one recorded specimen (FMNH PM 153, maxillary fragment with dp4; length = 13.9 mm) represents a juvenile individual substantially larger than any example of its late Du2 descendant, P. dunagani (mean dp4 length = 9.3 mm, observed range = 8.4–9.7 mm, sample size = 8), but one that appears to have been comparable in body size to middle early to latest Ch2 P. macrorhinus (includes middle early Ch2 Merycoidodon presidioensis; Lander 2019a:150) (see Wilson 1971a:46, table 31). As indicated by Wilson (1971a:46, fig. 28b), a small cusp similar to the one developed just beyond the posterolobial end of the short dp4 postprotocrista in FMNH PM 153 is found in the M1 of P. dunagani, too. Such cusps are also developed variously in the M1–3 of P. macrorhinus (including M. presidioensis type specimen, TMM 40505-2), Ch3 P. natronensis (includes Merycoidodon forsythae = Miniochoerus forsythae; Lander 1998:411), and even occur rarely in Or1–Or2 examples of P. periculum (e.g., F:AM 45109, 45238), as well, but are not recorded in dentally more primitive Aclistomycter middlettoni (E.B. Lander unpub. data). Prodesmatochoerus macrorhinus is considered herein to be of middle early to latest Ch2 (not Ch2–Ch3) age, whereas P. natronensis is strictly Ch3 in age (see below).

Prothero and Emry (1996:676) recommended that the (lower part of the) Chambers Tuff Fm that produced the (Lower) Porvenir LF in the Sierra Vieja (i.e., Big Red Horizon) be designated as the stratotype for Du2. The corresponding biozone could be named after the (lower part of the) Chambers Tuff Fm that produced Wilson (1971a:30–31) is first recorded less than 0.3 m (1 ft) above the base of the Big Red Horizon. No lower fossil occurrence appears to have been reported from the unit. The top of the stratotype is tentatively located within 1.5 m (5 ft) of the contact between the Big Red Horizon and the overlying Lower Marker Bed at TMM loc 40492 (see Wilson 1978:fig. 7, table 5). Aclistomycter (i.e., A. middlettoni) and Prodesmatochoerus n. sp. A (large) are restricted to the subage (Fig. 3) and, therefore, are specified herein as index taxa for middle early Du2.

**Middle early-late early Du2 boundary.** Comparatively few species recorded by Wilson (1978:table 6, fig. 7) have RFADs in the lower (but not lowermost) part or 26.8 mm (88 ft) of the Chambers Tuff Fm overlying the Lower Marker Bed in the Sierra Vieja. He identified the fossil-bearing interval as the Blue Cliff Horizon. The corresponding land mammal assemblage was regarded as late early Du2 in age and referred to as the Upper Porvenir LF by Lander (2019a:147, 150, 2019b:209). One species that also represents a RFAD but was not listed by Wilson (1978:table 6) is the cylindrodontid Dolocylindodon texanus (Fig. 3). However, it was listed by Wilson (1978:table 3, table 14, p. 27 as Pseudocylindodon texanus from the undivided Porvenir LF. Locality data for the one individual reported to be from the Porvenir LF at TMM loc 40636 (not 40646) by Wood (1974:39) indicates that it is from Riffle Range Hollow. Critically, TMM loc 40688, also in Riffle Range Hollow, was reported to be in the Blue Cliff Horizon by Gustafson (1986:19). No other fossil site appears to have been reported from Riffle Range Hollow. Therefore, the RFAD for D. texanus is probably a member of the Upper Porvenir LF from the Blue Cliff Horizon and late early Du2 in age (Fig. 3).

A second species not recorded by Wilson (1978:table 6) was the brontotheriid Duchesneodus uintensis. Locality data for specimens from the Chambers Tuff Fm reported by Mihlbachler (2008:316) indicate that they, too, are from the Blue Cliff Horizon (Fig. 3). The genus and species are based on individuals from the earliest (type) Du2 Lapoint Fauna (Fig. 3), which occurs in the Lapoint Mbr of the Duchesne River Fm in northeastern Utah. All but one of the 19 published specimens (i.e., type [CM 11809] and 17 topotype specimens) from the unit were recovered at the CM Titanotherium Quarry (Mihlbachler 2008:316), which was shown as lying on the eastern side of Twelvemile Wash by Kay (1934:plate 45). The quarry and the respective CM Titanotherium Quarry LF occur above the Lapoint Tuff, the basal unit of the Lapoint Mbr (see below). The one remaining or referred specimen (FMNH PM 22410) was found in Halfway Hollow. Because it is currently restricted to the earliest Du2 Lapoint Fauna,
the late early Du2 Upper Porvenir LF, and the correlative Tonque LF from the upper part of the Galisteo Fm in north-central New Mexico (Fig. 3). *D. uintensis* is designated herein as an index taxon for early Du2. The species was previously regarded as an index taxon for the Duchesnean NALMA, as recognized by Lucas et al. (2004b:118).

Locality data for the two FMNH individuals of *Quadratomus* *gigans* (FMNH PM 47–PM 48) reported from the Porvenir LF by Wood (1974:8) indicate that they are also from the Blue Cliff Horizon (Fig. 3). Mostly following Lander (2019a:147), late early Du2 is regarded in this report as that period in time during which (1) the generic RLADs of *Q.? gigans*, *Duchesneodus uintensis*, the rhinocerotid *Teletaceras mortvallisi*, and the leptomyercid *Hidrosotherium transpecosensis* (includes *Leptomeryx defordii*; see above), (2) the RLADs for the amphyconid *Daphoenictis n. sp. (small)* and the helaeltid *Colodon stovalli*, (3) the RFAD for *Dolocylindodon texanus*, and (4) the generic RFAD of the entelodontid *Cypretherium coarctatum* occurred together (Fig. 3; see Wilson 1978:fig. 7, table 6). Based on biostratigraphic and biochronologic data provided above, the Blue Cliff Horizon is designated as the stratotype for late early Du2 (= *late Viejan Subage*). Unfortunately, the lower boundary for late early Du2 cannot be defined stratigraphically with confidence because LLSSs or RFADs for age-diagnostic species are not reported accurately relative to their occurrences above the Lower Marker Bed. Therefore, the subage boundary is tentatively placed at the base of the Blue Cliff Horizon, which is designated as the late Du2 stratotype herein. The stratotype extends from 0–26.8 m (0–88 ft) above the Lower Marker Bed at TMM loc 40688 (see Wilson 1978:fig. 7, table 6). Along with *Duchesneodus uintensis*, *Colodon stovalli* is also regarded as an index species for early Du2, whereas *Quadratomus? gigans*, *Daphnoictis n. sp. (small)*, *Teletaceras mortvallisi*, and *Hidrosotherium transpecosensis* are identified as index species for middle early to late early Du2 (Fig. 3).

**Late early-late Du2 boundary.** Comparatively few taxa recorded by Wilson (1978:tables 9–10) from the Little Egypt LF in the uppermost part of the Chambers Tuff Fm represent RFADs. *Prodesmatochoerus dunaganii*, the smallest-bodied member of the genus, is found in the Reeves Bonebed at TMM loc 40209. The bonebed is 1.8 m (6 ft) thick and contains RFADs of species representing the Lower Little Egypt LF. It lies 12.2–14.0 m (40–46 ft) below the 4.6 m (15 ft) thick Salmon-Red Marker Bed and 16.8–18.6 m (55–61 ft) below the mutual tops of the marker bed and the Chambers Tuff Fm (see Wilson 1978:fig. 9). The Salmon-Red Marker Bed at TMM loc 40840 (= Chalk Gap Draw) contains species constituting the Upper Little Egypt LF, which includes the RLAD of *P. dunaganii*.

The (Lower) Little Egypt LF from the Reeves Bonebed was considered Ch1 in age by Prothero and Emry (1996:676, 2004:162) and a correlative of the (Lower) Rancho Gaitan LF from the Prietos Fm of northeastern Chihuahua (but latter assemblage actually correlative of middle early Du2 Lower Porvenir LF at Adobe Spring, based on shared occurrences of oreodontid *Aclistomycter middletonii*; Lander 2019b:209). Prothero and Emry (1996:676) stated that the Duchesnean-Chadronian NALMA boundary could be typified by the Reeves Bonebed, which they (Prothero and Emry 2004:162) designated as the Ch1 stratotype. It was suggested by Prothero and Emry (1996:676) that the (questionable) oreodontid *Bathygenys* (i.e., *B. reevesi*) as then recognized was the taxon best defining the beginning of the Chadronian NALMA because of its distinctiveness and its abundant occurrence during Ch1. Prothero and Emry (2004:162) mentioned *Bathygenys* Interval Zone as a possible name for the corresponding biozone. However, Lander (2013:56) assigned the late Ui3 agriochoerid *Protoreodon petersoni* to *Bathygenys* as its generic RFAD and most primitive member.

In addition to the “generic” RFAD for post-Ui3 *Bathygenys* (i.e., *B. reevesi*), Prothero and Emry (1996:676) suggested that the Duchesnean-Chadronian NALMA boundary could also be identified by generic RFADs of (1) the amphyconid *Brachyrhynchocyon* (i.e., *B. dodegi = Daphnoicyon dodegi*), (2) the entelodontid *Archaeotherium* (i.e., *A. cf. mortoni* of Wilson 1971b:12), and (3) the oreodontid *Merycodon* (i.e., *M. dunaganii = Prodesmatochoerus dunaganii*; see Lander 1998:412, loc SB44c). Except for *Archaeotherium cf. mortoni*, which occurred in the Salmon-Red Marker Bed at TMM loc 40840, the taxa were all found in the Reeves Bonebed at TMM loc 40209 (Fig. 3, Wilson 1971a:31, 40, 1971b:12, 1978:tables 9–10, Gustafson 1986:51). However, some biostratigraphic and biochronologic data they used were incorrect. For example, FMNH PM 98, which Wilson (1971b:12, 16, 1978:table 6) assigned to *A. cf. mortoni*, was reportedly from ca. 61.0 m (200 ft) above the base of the Chambers Tuff Fm. Locality data indicate that the specimen was actually found 1.5–15.2 m (5–50 ft) above the base of the Blue Cliff Horizon. Moreover, FMNH PM 98 might be referable to the entelodontid *Cypretherium coarctatum*, according to Foss (2007:125). Consequently, the generic RFAD of *C. coarctatum* really pertains to the late early Du2 Upper Porvenir LF (Fig. 3). Foss (2007:126) cited Wilson (1971b:13) when reporting a new species of *Cypretherium* from the Chambers Tuff Fm. However, page 13 was Wilson’s (1971b) figure 4, in which was illustrated an individual of *A. cf. mortoni* (TMM 40840–15) from the Upper Little Egypt LF of the Salmon Red Marker Bed. The three TM specimens from the marker bed are also assigned questionably to *C. coarctatum* in this report because they were found above a specimen that might be referable to *C. coarctatum* (Fig. 3) and preceded middle early Ch2 records of *C. cf. coarctatum* from the middle and upper parts of the Ahearn Mbr reported by Clark and Beerbower (1967:52). The Ahearn Mbr is the lowermost
unit of the Chadron Fm in the Chadronian NALMA type area of the Big Badlands in southwestern South Dakota (see Clark 1967:fig. 2).

Similarly, Lander (2019b:209) recorded the RFAD for *Bathygenys reevesi* from the Red Table LF, which occurs at TMM loc 41965 and underlies the generic RFAD of the middle early Du2 index taxon *Aclistomyceter middletoni* in the Montgomery Bonebed LF at TMM locs 41967–41968 (Fig. 3; Wilson 1986:355–356, table 7). The two assemblages occur in the middle part of Bandera Mesa Mbr of the Devils Graveyard Fm in Trans-Pecos Texas (Wilson 1986:355). Accordingly, the “generic” RFAD for *Bathygenys* as previously recognized by other authors (i.e., first post-U13 occurrence) predates the Little Egypt LF. Although the respective specimens are dentaries that cannot be assigned confidently to *B. reevesi*, they are about the same age as individuals of *B. reevesi* from the middle early Du2 Lower Rancho Gaitan LF that are represented by partial skulls with incomplete cheek tooth dentitions (Fig. 3, Lander 2019b:209; see Ferrusquia-Villafranca 1969:123, fig. 6b). The Lower Rancho Gaitan LF also contains *Aclistomyceter middletoni* (Lander 2019b:209). Despite occurring below *A. middletoni* of the Montgomery Bonebed LF, the Red Table LF is still regarded tentatively herein as middle early Du2 in age because, like the Lower Porvenir LF of the Big Red Horizon, it also contains the generic RFAD for *Agrochoerus* n. sp. (medium) at TMM locs 41965 and 42149, but not *A. middletoni* (Fig. 3; see above).

The RFAD for *Merycoidodon dunagani* (= *Prodesmatochoerus dunagani*) in the Lower Little Egypt LF no longer represents a generic RFAD, either (see above, Fig. 3). Critically, neither the unnamed, middle early Ch2, large-bodied species of *Prodesmatochoerus* (i.e., *P. n. sp. B* [large] of Lander 2019a:150, 2019b:209) nor the similarly large-bodied, middle early to latest Ch2 *P. macrorhinus* (includes *Merycoidodon presidioensis*) occurs in the Lower Egypt LF. *Prodesmatochoerus* n. sp. B (large) (CM 9391) is a member of the Lower Ahearn LF (new rank) of the Ahearn Fauna from the lower part (but not base) of the Ahearn Mbr. It represents the oldest Chadronian record of *Prodesmatochoerus*, whereas the Basal Ahearn LF (new; underlies Lower Ahearn LF) of the Ahearn Fauna is the oldest assemblage of definitive Chadronian (i.e., earliest Ch2) age in the Chadronian NALMA type area of southwestern South Dakota and adjacent northwestern Nebraska. Although the earliest to middle early Ahearn Mbr is the lowermost unit of the Chadron Fm in southwestern South Dakota and the overlying, latest Ch2 Peanut Peak Mbr is the uppermost unit, the Peanut Peak Mbr is, instead, the lower unit of the Chadron Fm and Chadronian NALMA stratotype in northwestern Nebraska (Terry 1998:figs. 7–9).

With the removal of *Archaeotherium* (i.e., *Cypretherium coarctatum*), *Bathygenys* (i.e., *B. reevesi*), and *Merycoidodon* (i.e., *Prodesmatochoerus dunagani*) from consideration as taxa defining the Duchesnean-Chadronian NALMA boundary, the only taxon left in Prothero and Emry’s (1996:676) compilation is *Brachyrhynchocestus dodgei* of the Lower Little Egypt LF from the Reeves Bonebed at TMM loc 40209. However, the presence of *P. dunagani* in the Lower Little Egypt LF instead of middle early Ch2 *P. n. sp. B* (large) or *P. macrorhinus* indicates that the assemblage is late Du2 rather than Ch1 in age (Fig. 3). Therefore, the generic RFAD of *Brachyrhynchocestus dodgei* is considered herein to represent a pre-type Chadronian record of the taxon (Fig. 3).

Lengths for lower cheek teeth of late Du2 *Colodon* n. sp. (small) (= C. n. sp. of Lander 2019a:150; TMM 40209-759, left dentary with p2–m3) of the Lower Little Egypt LF are less than those for early Du2 individuals of *C. stovalli* from the Lower and Upper Porvenir LFs and the Titus Canyon and Lapoint Faunas (E.B. Lander unpub. data, Rasmussen et al. 1999:424, Lander 2019a:150; see Stock 1949p. 236, Wilson and Schiebout 1984:table 3). Lengths for each of the p4–m1 in the specimen from the Lapoint Mbr cited by Lander (2019a:147, table 3) (CM 25745, dentary fragment with p4–m1) are 11.1 and ca. 14.0 mm, respectively (Lander unpublished data).

Based on biostratigraphic and biochronologic data presented above, the late early-late Du2 boundary is defined by (1) the RFADs of *Colodon* n. sp. (small) and *Prodesmatochoerus dunagani*, and (2) the generic RFAD of *Brachyrhynchocestus dodgei* in the Lower Little Egypt LF of the Reeves Bonebed at TMM loc 40209 (Fig. 3; see Wilson 1978:fig. 9). Accordingly, the late early-late Du2 boundary is situated at the base of the Reeves Bonebed or 18.6 m (61 ft) below the mutual tops of the Salmon-Red Marker Bed and the Chambers Tuff Fm. Similarly, the Upper Little Egypt LF, which occurs in the Salmon-Red Marker Bed at TMM loc 40840 and contains the RLAD of *P. dunagani* (see Wilson 1978:table 10), is regarded in this report as the youngest assemblage of late Du2 age. Therefore, the fossil-bearing interval spanning the upper 18.6 m (61 ft) of the Chambers Tuff Fm is designated in this report as the stratotype for late Du2 and the Little Egypt LF is identified as its type assemblage. Critically, Ch1 (= earliest Chadronian NALMA of Prothero and Emry 1996:676, 2004:162) is equivalent to late Du2 of this report and the latest Duchesnean NALMA of Lander (2019a:150). Consequently, Ch1 is rejected herein in favor of late Du2. The only assemblage confidently assigned to Ch1 by Prothero and Emry 1996:676), the Little Egypt LF, is clearly older than (1) the type earliest Ch2 Basal Ahearn LF from the basal part of the Ahearn Mbr of the Chadron Fm in the Chadronian NALMA type area of the Big Badlands in southwestern South Dakota and (2) its correlative, the Lower Yoder LF from the lower part of the exposed stratotype for the Yoder Mbr of the Chadron Fm at Goshen Hole in adjacent southeastern Wyoming, as discussed below. It is older still than the latest (type) Ch2 (not Ch4; see above) Peanut Peak Fauna from the Peanut Peak Mbr of the Chadron Fm in the Big Badlands.
The last assemblage contains the latest Ch2 RLAD of Prodesmatochoerus macrorhinus (includes Merycoidodon lewisi; E.B. Lander unpub. data), perhaps the largest-bodied member of the genus.

**Duchesnean-Chadronian NALMA (= Du2-Ch2) boundary.** The beginning of the Chadronian NALMA is marked only secondarily by the RFAD of the middle early Ch2 oreodontid Prodesmatochoerus n. sp. B (large) (= Merycoidodon sp. of Clark and Beerbower 1967:55; CM 9391, dentary fragments with m1–3). It occurs in the lower (but not basal) part of the Ahearn Mbr of the Chadron Fm in the Chadronian NALMA type area of the Big Badlands in southwestern South Dakota. The respective assemblage is the Lower Ahearn LF (new rank) of the Ahearn Fauna (see Lander 2019a:149). The earliest Ch2 RLAD of the small-bodied equid Mesophilus viejensis is found at the base of the Ahearn Mbr (Clark and Beerbower 1967:47) and presumably at a level below that of Prodesmatochoerus n. sp. B (large). The middle early Ch2 RFADs of larger-bodied species of Mesophilus, M. celer and M. hypostylus of Clark and Beerbower (1967:47), occur higher in the Ahearn Mbr (see Clark and Beerbower 1967:fig. 18.3 [P1–4 lengths for M. celer actually for M1–3], fig. 18.5, p. 39, but note inconsistencies regarding levels). Contrary to Lander (2019a:148), the assemblage from the basal part of the unit is regarded herein as distinct from the Lower Ahearn LF and is reassigned to the monotypic Basal Ahearn LF of the Ahearn Fauna. The RLAD of Mesophilus viejensis (includes ?M. sp. of Schlaikjer 1935:82 and M. texanus; Lander 2019a:146, 150; see Emry et al. 1987:134) also occurs in the correlative or earliest Ch2 Lower Yoder LF from the unit constituting the lower 4.6 m (15 ft) of the exposed stratotype for the Yoder Mbr of the Chadron Fm at Goshen Hole in southeastern Wyoming (Lander 2019a:150; see Schlaikjer 1935:fig. 2). However, the Lower Yoder LF also includes the generic RFAD for the leptomerycid “Leptomeryx” yoderi and the LLSO of the amphicyonid Brachyrhynchos cyanodon dodegi (= Miacis matthewi Schlaikjer 1935:77; Kihm 1987:34) (see Schlaikjer 1935:fig. 2). Following Wood et al. (1941:11, 37, plate 1) and contrary to Prothero and Emry (1996:476), the Lower Yoder LF and, by correlation, the Basal Ahearn LF are considered in this report to be of earliest instead of late early Chadronian age. Accordingly, the Duchesnean-Chadronian NALMA boundary is defined by the generic RFAD of “Leptomeryx” yoderi at an unspecified level in the lower 4.6 m (15 ft) of the stratotype for the Yoder Mbr.

**Preliminary subdivisions of Chadronian NALMA.**

The following subdivisions or subages of the Chadronian NALMA are based on those proposed by Prothero and Emry (1996:676–677), but have been substantially modified herein. They are based primarily on detailed biostratigraphic analysis of sections (1) at McCarty’s Mountain in southwestern Montana (CM Beds R–Z in Bone Basin Mbr of Renova Fm), (2) at Flagstaff Rim in central Wyoming (White River Fm, particularly with regard to Ashes B–G), and (3) in the Big Badlands of southwestern South Dakota (Ahearn and Peanut Peak Mbrs of Chadron Fm). Ch1 was rejected above. The following subages are tentative and subject to change.

1. Ch1: Rejected above. Equivalent to late Du2 of this report.
3. Middle early Ch2: Biochrons of agriochoerids Eomeryx minimus (generic RLAD), Proteodon n. sp. C (generic RLAD), oreodontids Leptauchenia platycetes (generic RFAD), Oreonetes anceps (generic RFAD), Prodesmatochoerus n. sp. B (large), P. n. sp. C (medium). RFADs of equid Mesophilus celer/hypostylus, oreodontid Bathygynus alpha, oreodontid Prodesmatochoerus macrorhinus. Probable RLAD of agriochoerid Agriochoerus n. sp. A (medium).
4. Late early Ch2: Biochrons of oreodontids Leptauchenia n. sp. (large), Oreonetes anceps douglassi (specific RLAD).
7. Ch3: Biochrons of secondarily smaller-bodied species of Bathygynus (i.e., B. hedlundae; generic RLAD), Prodesmatochoerus (i.e., natriensis; includes Merycoidodon/Miniochoerus forsythae). Biochron of Oreonetes n. sp. (small) (= RFAD of O. gracilis species group and structural grade). RLAD of Agriochoerus n. sp. C (medium).

**Biostratigraphy and biochronology of monotypic Protitanops**

The braontotheriid Protitanops curryi occurs in several assemblages of Du2 age. The Chambers Tuff Fm of Trans-Pecos Texas contains three stratigraphically superposed LF s, each of which includes P. curryi (Fig. 1, site 2, Fig. 3, Mihlbacher and Prothero 1921:7; see above). The middle early Du2 Lower Porvenir LF occurs in the lowermost part or Big Red Horizon (i.e., below Lower Marker Bed) of the Chambers Tuff Fm (see Wilson 1978:fig. 7, table 4). It includes the generic RFAD for P. curryi. The late early Du2 Upper Porvenir LF occurs in the Blue Cliff Horizon (i.e., above Lower Marker Bed) (see Wilson 1978:fig. 7) and also contains P. curryi. The generic RLAD of P. curryi from TMM loc 40804 (i.e., Chalk Gap Draw) occurs in the late Du2 Upper Little Egypt LF of the Salmon-Red Marker Bed at the top of the Chambers Tuff Fm (Wilson 1977:8, 10, 1978:fig. 9, 10). Additional records of P. curryi were found in the lower or middle part of the Chisos Fm at TMM locs 40932 and 41916 in Big Bend National Park, Trans-Pecos
Texas (Wilson 1977:10, fig. 5, Mihlbachler and Prothero 1921:7, 18). Specimens from the Salmon-Red Marker Bed at TMM loc 40804 and TMM 40932-1 from the Chisos Fm were originally assigned to *Menodus* by Wilson (1977:8, 10, fig. 5).

The Lower and Upper Porvenir LFs share a number of age-diagnostic taxa with correlative land mammal assemblages. They include (1) the Titus Canyon Fauna from the Titus Canyon Fm in the Grapevine Mountains in Death Valley National Park, California, and (2) the Rancho Gaitan LF from the Prietos Fm in the Cerros Prietos area ca. 32 km (20 mi) northwest of Ojinaga in northeastern Chihuahua, Mexico (Fig. 1, sites 1, 3, respectively, Mihlbachler and Prothero 1921:7). The Titus Canyon Fauna occurs in that part of the Lower Red Beds underlying the CIT Loc 255 Tuff Bed, which lies between the first and second red beds below the top of the Lower Red Beds (Lander 2019a:149, fig. 3, section in ”western fork of Titus Canyon”).

*Protitanops curryi* is based on a dentally small individual (CIT 1854, skull and mandible; Fig. 2a, Stock 1936a:656, plate 1, Mihlbachler 2008:298, figs. 147–148, Mihlbachler and Prothero 2021:fig. 3a) from the Titus Canyon Fm. It constitutes the monotypic Lower Titanotherium Canyon LF, which occurs at CIT Loc 253 in upper Titanotherium Canyon (Fig. 3, Stock 1936a:656, Lander 2019a:fig. 2, tables 2–3). CIT Loc 253 is situated near the bottom of the Lower Red Beds and well below the CIT Loc 255 Tuff Bed (Lander 2019a:fig. 3, section in ”upper Titanotherium Canyon”). *Protitanops curryi* also occurs in the late early Du2 Upper Porvenir LF, which contains (1) the generic RLAD for the ischyromyid *Quadratodus*? *gigans* and (2) the RLAD for the amphiomyid *Daphoenictis* n. sp. (small), as well (Fig. 3, Lander 2019a:table 3). On the other hand, *Daphoenictis* n. sp. (small) is associated with *Quadratodus*? *gigans* in the East Fork of Titus Canyon LF, which was found in a thin sandstone bed lying only a short distance above the base of the Lower Red Beds at CIT Loc 257 (Fig. 3, Lander 2019a:147, tables 2–3, fig. 3, section in ”eastern fork of Titus Canyon”). *Daphoenictis* n. sp. (small) might also be associated with *Dolocylindrodon texanus* in one of two fossil-bearing red beds containing the West Fork of Titus Canyon—Southwest LF at CIT Loc 254 (Fig. 3, Lander 2019a:147, tables 2–3). However, CIT Loc 254 comprises three sites (Lander 2019a:147), including (1) a quarry (Quarry 1 herein) in a thin dark reddish-brown red bed immediately above an even thinner light yellowish-gray tuffaceous sandstone bed at the base of the Lower Red Beds and (2) two other quarries (Quarries 2–3 herein) situated in a thick, bright reddish-orange red bed lying only two red beds higher in the local succession (E.B. Lander unpub. data). The West Fork of Titus Canyon—Southwest LF at CIT Loc 256 appears to be from the same thick reddish-orange red bed as Quarries 2–3 (E.B. Lander unpub. data). Consequently, the assemblages from CIT locs 253–254 and 256–257 represent the lowest and oldest assemblages constituting the Titus Canyon Fauna. Critically, *Quadratodus*? *gigans*, *Dolocylindrodon texanus*, and *Daphoenictis* n. sp. (small) are associated only in the Titus Canyon Fauna and the Upper Porvenir LF. Consequently, the Titus Canyon Fauna is no longer regarded as partly coeval with the Lower Porvenir LF (i.e., it is not also of middle early Du2 age), contrary to Lander (2019a:147, 150, 2019b:210). Instead, the entire Titus Canyon Fauna is considered herein to be early Du2 in age and a correlative of just the Upper Porvenir LF (Fig. 3). There is currently no information supporting correlation of the ”Lower” Titus Canyon Fauna and the Lower Porvenir LF.

A dentally large individual possibly assignable to *Protitanops curryi* (CIT 2007, partial skull; Stock 1936a:661, plate 2, Mihlbachler 2008:302, Mihlbachler and Prothero 2021:18, fig. 3b) is a member of the Upper West Fork of Titus Canyon—Northeast LF, which occurs in the first red bed below the CIT Loc 255 Tuff Bed at CIT Loc 255 and is one of the two highest and youngest assemblages constituting the Titus Canyon Fauna (Lander 2019a:149–150, fig. 3, section in ”western fork of Titus Canyon”). The specimen was previously identified as a menodontine? bronotheriid, cf. *P. curryi*, and *Protitanops* n. sp. by Stock (1936a:660, plate 2), Mihlbachler (2008:appendix 1, p. 435, etc.), and Lander (2019a:150, tables 2–3), respectively. Although CIT 2007 is too incomplete to allow definitive taxonomic assignment and represents an unusually large individual of *Protitanops*, there is no character of the skull or dentition that excludes the specimen from *P. curryi* (Mihlbachler 2008:320, Mihlbachler and Prothero 2021:18). The one preserved frontonasal horn (currently misplaced), like those in the type specimen, is bulbous and oval in cross section. The Upper West Fork of Titus Canyon—Northeast LF also contains the early Du2 heletaletid index species *Colodon stovalli* (see above), which, like *P. curryi*, is found in the Big Red and Blue Cliff Horizons of the Chambers Tuff Fm and, therefore, is a member of the Lower and Upper Porvenir LFs, as well (Fig. 3, Wilson and Schiebout 1984:10, 14, Lander 2019a:150, tables 2–3). According to Lander (2019a:150), the RLAD of *C. stovalli* in the Upper Porvenir LF is succeeded by smaller-bodied C. n. sp. (small), which is found in the late Du2 Lower Little Egypt LF from the Reeves Bonebed (Fig. 3; see above). Based on biostratigraphic data presented above, the HILSO of *C. stovalli* in the Upper Porvenir LF and its presence in the assemblage from CIT loc 255 suggest that (1) both records represent the species’ RLAD and, therefore, (2) the Titus Canyon Fauna is no younger than late early Du2 in age (Fig. 3). Consequently, the Titus Canyon Fauna is now considered to be entirely of late early Du2 in age, contrary to Lander (2019:150, tables 2–3). Similarly, the RLAD of *Protitanops curryi* in the late Du2 Upper Little Egypt LF of the Salmon-Red Marker Bed at TMM locality 40840 (= Chalk Gap Draw) also represents the RLAD for
monotypic Protitanops (Fig. 3), which is designated herein as a middle early to late Du2 index taxon.

A second dentally small individual of P. curryi was previously assigned to Protitops cf. B. brachycephalus by Ferrusquia-Villafranca (1969:106, figs. 3–4) (Fig. 3, Mihlbachler and Prothero 2021:4, 7). It was found at the base of the Upper Conglomerate Mbr of the Prietos Fm at loc 1 of Ferrusquia-Villafranca (1969:105, fig. 2, plate 1). That record constitutes the monotypic Upper Rancho Gaitan LF. The Lower Rancho Gaitan LF occurs 38 m (125 ft) below the top of the underlying Upper Tuff Mbr at locs 2–3 of Ferrusquia-Villafranca (1969:112, fig. 2, plate 1). It contains the middle early Du2 oreodontid index species Aclistomycter middletoni (Fig. 3, Lander 2019b:209). Therefore, the Upper Rancho Gaitan LF is no older than middle early Du2 in age, probably no younger than late early Du2, and (3) clearly no younger than late Du2.

**Records excluded from Protitanops curryi**

Numerous specimens from the Chambers Tuff Fm and the lower part of the Chisos Fm in Trans-Pecos Texas were previously assigned to the brontotheriid Menodus bakeri by Wilson (1977:8, 10) (Fig. 1, site 4). However, only some have been reassigned to Protitanops curryi by Mihlbachler and Prothero (2021:7), who rejected M. bakeri, regarding it as was a nomen dubium because its type specimen, a mandible, was taxonomically indeterminate (Mihlbachler and Prothero 2021:21–22).

Additional dentaries considered possibly referable to P. curryi were recorded from units of supposed Chadronian age by Swedler et al. (2021:48) and Lucas et al. (2004b:116–117). Those units included the Antero Fm of central Colorado and the Big Sand Draw Sandstone Lentil of the White River Fm in central Wyoming, respectively. The assemblage from the Big Sand Draw Lentil was regarded as early Chadronian in age by Emry et al. (1987:133–134) and Lucas et al. (2004b:117–118), based partly on the presence of (1) an equid (USNM 19108) comparable in body size to Mesohippus texanus (= M. viejensis; Mason 1988:209, Lander 2019a:146), (2) a dentally primitive, small-bodied rhinocerotoid (USNM 52143–521244) assigned to the hyracodontid cf. Hyracodon sp. or the rhinocerotid Caenopus yoderensis (nomen dubium; Prothero 2005:34), and (3) the entelodontid Brachyhyops wyomingensis (CM 12048, type specimen). The m3 length for the Mesohippus specimen (12.3 mm; Emry et al. 1987:134) is entirely compatible with those for individuals of M. viejensis from the middle early to late early Du2 Porvenir LF from the lower third of the Chambers Tuff in Trans-Pecos Texas (see Clark and Beerbower 1967:fig. 18.2). Therefore, USNM 19108 is assigned herein to that species. Similarly, lengths provided by Lucas et al. (2004b:117–118) for each of the P2–3 and p2–m3 of the two rhinocerotid specimens are comparable to those provided by Prothero (2005:table 4.1) for examples of the rhinocerotid Penetrigoniinae.

Consequently, USNM 52143–521244 is identified in this report as P. hudsoni. Elsewhere, however, Protitanops curryi, Mesohippus viejensis, Penetrigonia hudsoni, and Brachyhyops wyomingensis are all associated only in the middle early Du2 Lower Porvenir LF from the Big Red Horizon of the Chambers Tuff Fm (Fig. 3). Consequently, the assemblage from the Big Sand Draw Sandstone Lentil is considered tentatively herein to be of middle early Du2 instead of early Chadronian age.

Unfortunately, the brontotheriid specimens from the Antero Fm and the Big Sand Draw Sandstone Lentil all lack diagnostic features allowing their identification even to the generic level. Accordingly, their identifications as Protitanops curryi are only tentative (Mihlbachler and Prothero 2021:24–25). Moreover, the example from the Antero Fm is younger than any definitive record of the genus. Du1 records in the Hancock Mammal Quarry LF from Unit E of the Clarno Fm in north-central Oregon were previously identified as Protitanops n. sp. and P. curryi by Hanson (1996:232) and Lucas et al. (2004a:91), respectively. They have since been reassigned to Eubrontotherium clarinoensis by Mihlbachler (2007:20).

**Geochronology**

**Vieja Group, Texas.** The biochron of monotypic Protitanops (i.e., P. curryi) is constrained by SLCLF Ar/Ar ages for the Buckshot Ignimbrite and overlying Bracks Rhylolite, which bracket (1) the Chambers Tuff Fm of Sierra Vieja in Trans-Pecos Texas, (2) the middle early to late Du2 Porvenir and succeeding Little Egypt LFs, and (3) the RFAD and RLAD of Protitanops (Fig. 3; see Wilson 1978:fig. 5). The most recently published date for the Buckshot Ignimbrite, 37.68 ± 0.04 Ma, was calculated using the 28.201 Ma date for FCTs (Christopher D. Henry 2011 written communication in Kelly et al. 2012:235). The age of 36.67 ± 0.08 Ma provided by Prothero and Swisher (1992:table 2.2, p. 69, footnote, p. 73) for the Bracks Rhylolite was calculated using the previously accepted date of 520.4 Ma for the Mmb-1 flux monitor. That date is recalculated herein to 37.15 ± 0.08 Ma, as prescribed above. These two current dates indicate that the Protitanops biochron and middle early to late Du2 are earliest late Eocene in age, as indicated by the Geologic Time Scale 2020 (Speijer et al. 2020:fig. 28.12, p. 1114).

**Duchesne River Fm, Utah.** Kay (1934:366), “lithologic cross-section of the Uinta and Duchesne River Formations,” plates 45–46) provided bed-by-bed lithostratigraphic descriptions (“lithologic cross-section”), a stratigraphic columnar section, and a geologic cross section for the Lapoint Horizon of the Duchesne River Fm in the Uinta Basin of northeastern Utah. He compiled the corresponding data in 1929 (Peterson and Kay 1931:301, Kay 1934:359, footnote 4), along the eastern side of Twelvemile Wash and north of Vernal-Lapoint Road (= Utah State Route 121) in the Uinta Basin. The CM Titanotherium Quarry occurred in cross-section unit 226, which represented a 0.9 m (3 ft) thick bed of fine-grained
sandstone and conglomerate. A hand-written annotation presumably authored by J. LeRoy Kay occurs at the right-hand margin of the cross section on page 366 and adjacent to the description for underlying unit 219. The entry is in a bound copy of Volume 23 of the *Annals of the Carnegie Museum* that is archived at the CM. Critically, the annotation indicated that unit 219, a bluish-white clay layer 6.7 m (22 ft) in thickness, represented the “base of [the] Lapoint” Horizon (i.e., lower contact of unit 219 = mutual lower contacts of Lapoint Horizon and Lapoint Mbr). The Lapoint Tuff Bed is considered herein to be the same as (1) unit 219 of Kay (1934:366), (2) the lowermost extensive fine-grained bentonitic layer or unit 1, a 6.25 m (20.5 ft) thick, greenish-gray silty claystone bed, the base of which was equated with the lower contact of the Lapoint Mbr by Andersen and Picard (1972:14, appendix, “Lapoint Member,” p. 28), (3) the first or lowest prominent and laterally extensive or continuous ash bed of the Lapoint Mbr, with the lower contact of the ash bed being the same as that for the mbr, as reported and clearly shown by Webb (2017:11–12, fig. 8, table 2), and (4) the prominent 5.5 m (18 ft) thick tuff bed that Jensen et al. (2020:cover, p. i, explanation, figs. 2b, 3a, appendix 1, p. 1–1, samples DRF-E–DRF-G) reported and definitely showed as occurring below the level of Titanotherium Quarry and at the base of the Lapoint Mbr or at the contact between that mbr and the underlying Dry Gulch Creek Mbr (but shown incorrectly lying almost entirely at top of Dry Gulch Creek Mbr and extending only minimally into basal part of Lapoint Mbr in their fig. 2b). Although Murphey et al. (2011:160, fig. 15c, 2017:46, fig. 17c) stated that the upper and thicker of two bentonite beds underlying Titanotherium Quarry was the one identified as the La Point (= Lapoint) Tuff (Bed) by Prothero and Swisher (1992:49), they incorrectly considered the lower thinner tuff bed, instead, to be the one occurring at the Dry Gulch Creek-Lapoint Mbr contact. Kelly et al. (2012:232) also reported that the Lapoint Tuff Bed lay near rather than at the base of the Lapoint Mbr. Webb (2017:plate 1) appears to have included strata in the lowermost part of the Lapoint Mbr that Sprinkel (2007:plate 1) had earlier included in the uppermost part of the Dry Gulch Creek Mbr. However, Webb (2017:fig. 8) clearly used the Lapoint Tuff Bed as the basal unit of the Lapoint Mbr. Presumably, therefore, Sprinkel (2007:plate 1) used a tuff bed shown by Jensen et al. (2020:fig. 2b) overlying the 5.5 m thick tuff bed (= Lapoint Tuff Bed) as the basal unit of the Lapoint Mbr.

Prothero and Swisher (1992:49, table 2.2, p. 69, footnote, p. 73) provided a SCLF Ar/Ar date of 39.74 ± 0.17 Ma for the Lapoint Tuff Bed, which reportedly occurred at the contact between the Dry Gulch Creek and overlying Lapoint Mbrs. That date was also calculated using MMhb-1 and was subsequently recalculated to 40.26 ± 0.08 Ma by Kelly et al. (2012: 232). Webb (2017:11–12, 32–33, table 4, p. 65, fig. 3) and Jensen et al. (2020:6, 8, figs. 2b, 4b, table 1, table 3) but incorrectly identified as Dry Gulch Creek Mbr], appendix 1, p. 1-1) reported a SCLF Ar/Ar date of 39.47 ± 0.08 Ma for a second tuff bed in the Lapoint Mbr. The dated sample (DRF-A) was from a 0.5 or 1–2 m (ca 1.6 or 3.3–6.6 ft) thick tuff bed in the lower part of the mbr. That tuff bed overlies the CM Titanotherium Quarry level, but might still be bracketed by the Lapoint Fauna (Fig. 3; see above).

Webb (2017:11–12, 32–33, table 4, p. 65, fig. 3) and Jensen et al. (2020:6, 8, figs. 2b, 4a, table 1, table 3 [but incorrectly identified as Lapoint Mbr], appendix 1, pp. 1–1–1–2) recorded a SCLF Ar/Ar date of 39.36 ± 0.08 Ma for a third tuff bed in the Duchesne River Fm. The dated sample (DRF-H) was from a 0.25 m (0.8 ft) thick tuff bed lying 2–3 m (0.6–0.9 ft) below the contact between the Lapoint and underlying Dry Gulch Creek Mbrs (Fig. 3). That tuff bed was shown in their figure 2b lying at about the same level as the USNM Halfway Hollow Quarry, which produced land mammal remains previously attributed to the Lapoint Mbr by Emry (1981:563, 565) (Fig. 3; see above). Sample DRF-I came from the same tuff bed as sample DRF-H. The third tuff bed might represent the lower of two tuff beds underlying Titanotherium Quarry and the one that Murphey et al. (2011:160, fig. 15c, 2017:46, fig. 17c) identified incorrectly as the basal unit of the Lapoint Mbr or the contact between the Dry Gulch Creek and Lapoint Mbrs. The three dates for the Duchesne River Fm are consistent with a late middle Eocene age for the Lapoint and Halfway Faunas.

**Chadron Fm, South Dakota.** K-Ar analyses have yielded three dates for the Ahearn Ash Bed (new), which lies 0.9 m (3 ft) below the top of the Ahearn Mbr of the Chadron Fm in the Big Badlands of southwestern South Dakota (McDowell et al. 1973:11, Morris F. Skinner and Richard H. Tedford personal communications in Emry et al. 1987:148–149, note 15). The arithmetic mean for the three corrected dates is 36.8 Ma. That determination is consistent with (1) an early late Eocene or “Ch1” age for the Lapoint Tuff, (2) the SCLF Ar/Ar age of 37.15 ± 0.08 Ma for the Bracks Rhyolite, and (3) biostratigraphic and biochronologic data for the Basal Ahearn and Little Egypt LFrs (Fig. 3; see above). The two dates and lithostratigraphic and biostratigraphic data suggest the Duchesnean-Chadronian NALMA boundary is much closer to 37.15 Ma in age than it is to 36.8 Ma.

**Summary and Conclusions.**

1. The Uintan–Duchesnean NALMA (i.e., Ui3-Du1) boundary and the beginning of Du1 are defined by the RFAD for the medium-sized agriochoeroid *Diplobunops crassus*. The species’ LLSO is at an undetermined level above the brown sandstone bed constituting the basal 3.0–6.1 m (10–20 ft) of the Brennan Basin Mbr, the lowermost unit of the Duchesne River Fm and the Duchesnean NALMA stratotype in northeastern Utah. The corresponding Upper Randlett Fauna is the Du1 type assemblage.
2. The Du1-Du2 boundary is tentatively defined by the RFAD of *Epihippus intermedius*. The species’ LLSO is probably at a level in the Dry Gulch Creek Mbr of the Duchesne River Fm below those of SDSNH locs 5939, 6087, and 6339. That record of *E. intermedius* represents the LLSO of the earliest Du2 Halfway Fauna. The RFAD of the agriochoerid *Diplobunops* n. sp. (very large) from the uppermost part of the Dry Gulch Creek Mbr at Halfway Hollow Quarry represents the HLSO of the fauna.

3. The earliest Du2 Lapoint Fauna is from the Duchesne River Fm’s Lapoint Mbr, the uppermost unit of the Duchesnean NALMA stratotype. The CM Titanotherium Quarry LF is the type earliest Du2 or Lapointean assemblage. The quarry is over lain by a tuff bed determined to be 39.47 ± 0.08 Ma old.

4. The earliest-middle early Du2 boundary is tentatively defined by the generic RFAD of the agriochoerid *Agrioceras* n. sp. A (medium). The species’ LLSO is less than 0.3 m (1 ft) above the base of the Big Red Horizon, the lowermost unit of the Chambers Tuff Fm in Trans-Pecos, Texas. The middle early Du2 stratotype extends from that level to one perhaps within 1.5 m (5 ft) of the horizon’s top. The respective Lower Porvenir LF is the type middle early Du2 or early Viejan assemblage. It also contains the RFAD for the monotypic brontotheriid *Protitanops* (i.e., *P. curryi*).

5. The Upper Porvenir LF is the type late early Du2 or late Viejan assemblage. The corresponding stratotype extends from 0–26.8 m (0–88 ft) above the base of the Chambers Tuff Fm’s Blue Cliff Horizon. The assemblage includes the generic RFAD for the ischyromyid *Quadratomus? gigas* and the RFAD for the cylindroodontid *Dolocylindrodon texanus*. Late early Du2 is defined partly as that period of time during which the two species occurred together.

6. Based on shared occurrences of *Q.? gigas* and *D. texanus*, the Upper Porvenir LF is regarded as a correlative of the “Lower” Titus Canyon Fauna from the lower part of the Titus Canyon Fm’s Lower Red Beds.

7. The Little Egypt LF occurs in the Reeves Bonebed and overlying Salmon-Red Marker Bed in the upper 16.8–18.6 m (55–61 ft) of the Chambers Tuff Fm. The Lower Little Egypt LF and the generic RFAD of the amphichyrid *Brachyrhynchocyon dodgei* are found in the bonebed, whereas the Upper Little Egypt LF and the generic RFAD of *Protitanops curryi* are from the marker bed. The oreodontid *Prodesmatochoerus dunagani*, the smallest-bodied member of its genus, is restricted to the two units. Previously considered to be the Ch1 type assemblage, the Little Egypt LF lacks Ch2 *Prodesmatochoerus* n. sp. B (large) and large-bodied *P. macrorhinus*, which are among the largest-bodied records of the genus. The latter two species have RLADs in the Chadron Fm of the Chadronian NALMA type area in southwestern South Dakota. In contrast, *P. dunagani* is not recorded from the Chadron Fm or its correlatives. Therefore, the Little Egypt LF is regarded herein as the type assemblage for late Du2 instead of Ch1. Accordingly, Ch1 is considered to be equivalent to late Du2 and the oldest Chadronian assemblages are of earliest Ch2 age.

8. The Chambers Tuff Fm, the Porvenir and Little Egypt LF s and, by correlation, the Titus Canyon Fauna, along with the *Protitanops curryi* biochron and middle early to late Du2 are constrained by Ar/Ar dates of 37.68 ± 0.04 and 37.15 ± 0.08 Ma for the Buckshot Ignimbrite and Bracks Rhyolite, respectively. The two tuffs bracket the Chambers Tuff Fm. Therefore, the *P. curryi* biochron is ca. 0.53 Ma long. Critically, the Porvenir LF is not a correlative of the Lapoint Fauna, contrary to some workers beginning with Wilson (1971a:fig. 6, 1986:fig. 7). Instead, the Porvenir LF is about 1.79–2.32 Ma younger.

9. The Duchesnean-Chadronian NALMA (i.e., Du2-Ch2) boundary is defined by the earliest Ch2 RLAD for the equid *Mesohippus viejensis* and generic RFAD of the leptomerycid “*Leptomeryx*” *yoderi* in the Lower Yoder LF from unspecified levels in the lower 4.6 m (15 ft) of the stratotype for the Yoder Mbr of the Chadron Fm in southeastern Wyoming. The correlative Basal Ahearn LF, which also contain the RLAD of *M. viejensis*, is from the basal part of the Ahearn Mbr of the Chadron Fm in the Chadronian NALMA type area. Both assemblages represent the oldest assemblages of Chadronian (i.e., earliest Ch2) age.

10. The Ahearn Ash Bed, which lies 0.9 m (3 ft) below the top of the Ahearn Mbr, is 36.8 Ma in age, based on K-Ar analysis. That date and the Ar/Ar date of 37.15 Ma for the Bracks Rhyolite constrain the age of the Duchesnean-Chadronian NALMA boundary, which is probably much closer to 37.15 Ma in age than it is to 36.8 Ma.

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**Literature cited**


Recently formed microhabitat wetlands along the Coachella Canal, Imperial County, California

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ABSTRACT—The emergence and growth of twelve small wetlands next to and downgradient from the western embankment of the Coachella Canal is reported here. They were not present before the adjacent section of the canal was relocated and lined. Rapid wetland development and accompanying plant growth have attracted many animals. The origin of the wetland’s water is uncertain but is definitely associated with the canal. The sudden appearance of the wetlands and response by plants and animals demonstrates how quickly a water source transforms the desert.

1. Introduction

Microhabitats are small, localized environments whose biota is significantly different from that found in their more expansive immediate surroundings. Examples of microhabitats are decaying logs, shaded areas, coppice dunes, oases, road berms, north sides of trees, termite mounds, and caves. Even a dripping faucet produces a tiny bloom of life. Microhabitats occur all over the world and in all climates, and are often associated with local water sources. Most are natural. Others are manmade, like gardens, parks, terrariums, and agricultural fields. Microhabitats may be transient or enduring. Even within a single microhabitat, there may be smaller ones attributable to abiotic condition such as shade and soil variations.

Left to themselves, deserts change very slowly. But when previously dry areas suddenly turn green, it’s a good bet that a new water source is involved. During unrelated field work near the Coachella Canal (CC) in the Sonora Desert of Imperial County, California (Figure 1), I noticed a number of “suspiciously green” areas adjacent to the western embankment of a ~9 km section of the canal. The purpose of his paper is to bring these microhabitats (“wetlands”) to the attention of a broader audience and to report preliminary investigations of them.

2. The wetlands

The Coachella Canal carries Colorado River water from the All American Canal in Imperial County, California to the Coachella Valley of Riverside County. Along the way, some of the water is diverted for agricultural use in both counties. The canal is owned by the U.S. Department of the Interior, Bureau of Reclamation and operated by the Coachella Valley Water District (CVWD). Originally the canal was unlined and lost significant water to percolation into the soil. A new CC was built immediately west of the old canal and lined with a 30 mil polyvinyl chloride (PVC) membrane on the soil, and covered with 3 inches of concrete. It is bounded on both sides by embankments straddling the flowing water and reaching 3-10 m above grade. The flanks of the embankments slope about 30° and the top of each embankment is 6-8 m wide, flat and horizontal.

The small (~100 m), heavily vegetated wetlands discussed in this paper occur adjacent to the western downgradient embankment of a ~9 km section of the lined canal (Figures 1 & 2). I numbered them from south to north as W1, W2, W3, etc. Subsequent study using Google Earth revealed at least twelve such wetlands (Figure 3) that were easily identified by their enhanced vegetation and in some cases shallow, standing water. Each wetland was contiguous with the western embankment of the CC and first appeared during or after that section of the CC was relocated and then lined during the period 2004–2007 (Figures 4 & 5). Field visits revealed that most wetlands have associated dark, wet soil on the lower embankment flank and some occur higher than the canal water level. At the present time, moisture and water appear on soil mounds between the road and the western embankment. Standing water is also found in the former unlined canal bed.
All wetlands extend from the foot of the canal’s western embankment and across the Coachella Canal road where they extend downhill along pre-existing drainages by typically one hundred meters. Most drainages crossed by the canal showed no enhanced vegetation.

From the Coachella Canal road, the wetlands are obvious as isolated areas of dense green growth surrounded by otherwise tan, relatively barren badlands. They are along a section of the canal that runs perpendicular to the local surface stream and runoff flow. During the wet season, water ponds up against the raised eastern shoulders of the Coachella Canal road (dirt) and often flows across it at low points in the road to the drainage that continues to the west. Monthly visits during 2020 and 2021 revealed that standing water and/or wet mud is present year-round. Field examination showed significant amounts of muddy soil on the canal embankment above the wetlands. Salt cedar, honey mesquite and desert ironwood were the most common trees. The shallow standing water contains an assortment of plants, insects and algae and is often visited by birds and other animals. Based on ~20 years of Google Earth imagery, the wetlands are still growing.

Myriad erosion channels usually less than a meter wide extend from the top of the embankment and down its flank to local grade to the west. Some of the channels clearly originate from water runoff from the top of the embankments. Most of these channels are present along the entire length of the canal and most do not end in wetlands. In a few places the concrete panels are distorted and cracked. Repairs have been attempted on some (Figure 6).

Scattered along both sides of the canal are a number of wildlife drinkers. These are roughly
3. Discussion

Many of the observations reported in Section 2 are familiar and well understood in desert environments. Enhanced vegetation in any wetland is to be expected where surface water is found, especially in erosional drainages. Vegetation lineaments that are parallel to the eastern (upgradient) embankment of the former canal appear to be due to barriers to surface runoff where water ponds. These lineaments are common along the eastern embankment and their locations only partially correlate with the wetlands discussed here. Most vegetation lineaments perpendicular to and downgradient from the lined canal are persistent and are seen at many places, often predating canal lining. The height (i.e., age) of many of the trees in the wetlands that postdate the lined canal (up to 15 m), suggests a long-lived, enduring water source. Birds, insects, and larger mammals are attracted to standing water and vegetation, and their tracks were evident.

The exact cause of the wetlands is unclear but to identify the source of water, three aspects must be considered:

1. They are physically connected to the canal.
2. They appeared only after the lined canal was built.
3. Their water source is long-lasting, having persisted since ~2004.

Thus it seems obvious that the wetlands are fed by canal water in some way. Is this to be expected or is the canal leaking? As noted earlier, numerous cracks and buckles in the concrete lining were evident, as were attempts to patch the larger ones (Figure 6), so some leakage is occurring or has already happened. It is possible that cracks and slumps in the concrete have penetrated the underlying PVC membrane and caused leaks. Are they enough to account for the wetlands?

Other mechanisms may be influencing the wetlands. Many erosion channels on the western embankment flanks extend up to and partially incise the horizontal embankment top; the canal liner does not reach the top of the embankments so it would seem unlikely that CVWD would allow water to overtop the embankments. The channels were probably created by rainwater, which may contribute to some of the wetlands but not all: such channels are
4. Summary and conclusions

1. The wetlands are clearly associated with the canal. They are adjacent to the downgradient side of the canal and originated only after the canal was relocated and relined.

2. The wetlands are fed by a permanent though perhaps variable water source. They persist year after year and are still growing. The presence of evaporites suggests a persistent water source that is depositing salts at the surface as water evaporates.

3. The wetlands only occur in a small area 9 km long, and always in existing drainages. Similar vegetated regions rarely if ever form away from the canal.

4. The amount of wetland water varies seasonally but is usually present year-round.

In view of the possible importance of such microhabitats to desert biota and migratory birds, further investigation is warranted. Ecologically and hydrologically, the wetlands are interesting in themselves regardless of the processes that originated them.

Acknowledgements

I am indebted to Brian McNeece for assistance in the field. Andrea Donnellan drew my attention to wetland W10 and joined me during the initial visit to the area.

References

5. https://www.army.mil/article/247609/water_in_the_desert_wildlife_water_catchments_help_sustain_diverse_wildlife_at_u_s_army_yuma_proving_ground
7. Since the microhabitats are related to the canal, I contacted the CVWD but they did not reply to my calls and emails.
Old Dinah, abandoned in Daylight Pass, Death Valley, 1920s

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Abstract—Old Dinah is a steam-powered traction engine first used during the 1890s by “Borax” Smith under the Pacific Coast Borax Mining Company (PCB) at the colemanite mining town of Borate in Mule Canyon within the Calico Mountains of the Mojave Desert. An early photo of Old Dinah in Mule Canyon, from the U.S. Borax collection, appears in Hildebrand’s book (1982) and is dated 1894. This may be the year it was first used for hauling borate ore from the Calico Mountains.

The use of Old Dinah was an unsuccessful attempt to replace the 20 mule teams and wagons (20MT) for hauling colemanite ore and mining supplies, without the enormous expense of constructing a railroad. The narrow-gauge Borate & Daggett Railroad was eventually built to replace both the 20MT and Old Dinah, operating between the mines at Borate, the Marion Mill (near present-day Yermo) and the railhead at Daggett.

Old Dinah was placed into operation one last time during 1904 by Borax Smith as an experiment to check the feasibility of hauling refined borax, supplies and equipment between the closest railroad siding at Ivanpah in the Mojave Desert and the new PCB mine at the Lila C in the Amargosa Desert. The failure of the steam boiler on that run from Ivanpah towards the Lila C Mine gave Smith proof that another railroad, this time from the new mine, was the only answer to his most recent transportation dilemma. He began planning the Tonopah and Tidewater Railroad and Old Dinah became surplus and was later sold. After being purchased from PCB, Old Dinah was used in hauling mine supplies and ore to and from the Keane Wonder Mine. This is the story of why Old Dinah became abandoned in Daylight Pass during the 1920s.

Introduction

This article documents the history of Old Dinah and the personal quest in search of the abandonment site located approximately 100 meters southeast of Daylight Pass Road and northeast of Hells Gate within the dry wash of Boundary Canyon. In this remote region of present-day Death Valley National Park, Boundary Canyon forms the dividing line between the Grapevine Mountains to the northwest and the Amargosa Range towards the southeast.

This fieldwork was conducted by the author and Stephen Beck during November 2019. The search was based upon one black & white photo of Old Dinah taken circa 1928–1932 southwest of the geomorphic feature of Daylight Pass. Many successful searches by others have been conducted for the abandonment site of Old Dinah in this region of Death Valley. This article documents just one of the many successful field research endeavors.

Depletion of the mineral resource at Borate

After the year 1900, the borate deposits in the Calico Mountains, consisting primarily of colemanite, began showing signs of depletion after years of mining the limited mineral resource. Borax Smith instructed management of the Pacific Coast Borax Mining Company to begin focusing mining efforts on the rich colemanite deposit at the Lila C claims located on the northeastern edge of the Greenwater Range, located several miles southwest of present-day Death Valley Junction.

The finite life of a mine

Mining companies must first demonstrate to investors that a profit can be achieved when developing a naturally occurring mineral deposit. Factors that are essential to consider before mining can begin may include: ore quality, ore quantity and the marketability of the natural resource. All mines have a finite life. At times, sudden changes in the underground conditions within a mine can occur that affect the company’s ability to continue to operate at a profit.

These changes may include: the deposit(s) become depleted through the mining process, the mine begins running out of recoverable ore, the remaining ore in-place is of lower quality, the main ore vein is terminated by a fault, underground working conditions are too dangerous for the miners, groundwater infiltration is too expensive to pump, and the price/marketability of the ore has changed.

When mining conditions change drastically, then it may be time to close the mine and to move on to more favorable prospects. Throughout the history of mining in the West, a mine can be profitable one day and close its doors the following day. Many abrupt mine closures have had no early indications of the miners’ last day on the job.
Lila C mining claims

Borax Smith experimented with using Old Dinah to ship borate ore from the Lila C mining claims during 1904 (Old Dinah was manufactured by the Best Steam Traction Engine Company, circa 1894). The “Lila C” was named after William T. Coleman's daughter Lila. Coleman sold the Lila C claims, the Harmony Borax Works, the claims in the Calico Mountains and many other holdings to Borax Smith during the late 1880s. (The idle/abandoned Lila C mine is located several miles southwest of present-day Death Valley Junction.) In 1904, a traction road was graded between the railhead at Ivanpah and the mine at Lila C.

Having spent $100,000 on the traction road, Smith turned back to the Best steam tractor, hoping that this time it would work, given a proper roadbed. So Old Dinah was refurbished at Borate, shipped to Ivanpah by rail, greeted there by John Ryan, and put on the ground. Wagons were attached. After firing up, it started off with its load of empties trailing behind. Fourteen miles out, at State Line Pass, the engine blew out a boiler flue and the experiment was over. (Hildebrand, p. 73).

The failure of the steam traction engine resulted in Smith negotiating with Borax Consolidated Limited, PCB’s parent company, for the planning and construction of a standard gauge railroad that was later named the Tonopah & Tidewater Railroad (T & T). Planning began for the T & T, extending 121 miles from the Santa Fe Railroad siding at Ludlow to Death Valley Junction. From Death Valley Junction, a railroad spur extended to (Old) Ryan at the Lila C Mine. Mine production of colemanite from the Lila C claims began during June 1907.

Old Dinah is sold

During 1910, the steam traction engine was sold by Pacific Coast Borax Company to J. R. Lane of Calico. The boiler was rebuilt, and Old Dinah was placed back into service transporting ore and supplies between the Keane Wonder Mine in Death Valley, CA and the T & T siding at Gold Center near Beatty, NV. During the 1920s, after years of good service for J. R. Lane's operations, the boiler again failed in Daylight Pass while ascending the grade towards Rhyolite. The vertical design of the boiler and the use of high TDS (total dissolved solids) water led to scale buildup, which plugged steam tubes, and led to a ruptured pressure vessel.

Abandoned in Daylight Pass

Old Dinah was abandoned in Daylight Pass some time in the 1920s. The traction engine sat for many years at the site where the boiler blew out in Daylight Pass, a remote canyon on the east side of Death Valley. Iron parts were scavenged from the vehicle as it sat abandoned. During this time, the water tank was removed. In 1932, after years of sitting idle in Daylight Pass, Harry Gower of the Pacific Coast Borax Company used available company equipment to transport the engine approximately 35 miles to Furnace Creek in Death Valley. It was kept from the scrap pile by Gower’s insight. Since 1932, the traction engine remains at Furnace Creek and is on permanent exhibit in the National Park.

Transportation and early vehicles

The history of Old Dinah is just one example of the hardships associated with the task of providing transportation in the harsh desert environment of Death Valley and other desert regions of the West. Transportation was key to the success of any mine. Commodities that were essential to mining operations included supplies, timbers, explosives, equipment, water, food, shelter, and manpower. The extracted ore also had to be transported to a mill for processing, refining, calcining, and/or smelting.

The transition period between the use of wagons powered by draft animals and the trial periods of mechanical means of transportation were met with many failures. The excessive heat and the general lack of good roads and reliable supplies of fuel and water in the desert regions led to transportation disasters and many deadly consequences. A sequence of emerging technologies included steam-powered traction engines, railroads and gasoline-powered vehicles. However, each new vehicle design was often met with enormous costs, limited successes, mechanical problems and at times catastrophic failures.

Daylight Pass today

On November 9, 2019, Steve Beck and I were determined to find Old Dinah’s abandonment site in Daylight Pass. We only had one black & white photo to guide our search. The two photos in this article, old and new, indicate that we were within a few feet of the cameraman’s position when the old photo was taken circa the late 1920s. The profile of the mountains in the background and the lay-of-the-land in the foreground assisted us in locating the exact site. No evidence remains of Old Dinah at this location. The original dirt trail that the traction engine had been traveling on when the boiler blew was essentially a dry wash with intermittent seasonal rain runoff. All evidence of tracks and/or iron parts have been washed away or buried by over 90 years of time. The date of the black & white photo is unknown but is thought to be taken during the late 1920s.

Detailed analysis of the photos

After reviewing both photos, taken 90 years apart, I noticed that there were bushes within or near the lower left corners of the photos. After close examination, I theorized that these two occurrences of bushes could actually be the same plant still alive after a span of 90 years. The bush was identified during subsequent field
work in late November 2019 to be a creosote bush. Could the two bushes in the separate photos be the same organism? Yes, they could be. However, more research needs to be completed on the photos and on the creosote bush that is alive today.

**The creosote bush**

Current botanical research on the creosote plants have indicated that some “king clone” creosote rings are over 11,700 years old, having propagated new sprouts from an ancient living root bundle. Therefore, the age of some creosote plants corresponds in general age to the end of the last Ice Age, during a time when an abundance of surface water was still flowing through a chain or series of Pleistocene lakes that connected groundwater basins throughout the Mojave Desert and other desert regions of the Western States.

The original photo taken during the late 1920s does not exhibit the resolution needed to positively identify the plant. A slight wind on the day the early photo was taken could have easily blurred its image. However, botanical evidence involving carbon dating of dead plant material at the surface and within the soil could reveal an oldest age date. Excavating to the root bundle for a radiocarbon age date (Carbon-14) is too invasive for the plant. Any excavation conducted within a National Park can only be achieved with approval by the National Park Service.

**Desert history**

Desert history is full of stories that include the hardships that had to be overcome for common people to scratch out a living. The harsh desert conditions included: intense heat, discomforts, pain and at times...
death associated with travel across the great distances in remote lands. The general lack of water and the difficulties in growing crops and raising livestock in arid regions added to the hardships.

Mining in the desert during the early years
Our modern economy is based upon continuous successes in overcoming tremendous obstacles to produce a profit. Mining in the desert during those early years represents the best examples of achievements made under extraordinary circumstances to keep people employed and for a company to produce a monetary gain, no matter how insignificant.

Our strong economy of today is based upon the continuous sequence of accomplishments by common citizens who made tremendous sacrifices as they sought to feed their families and to earn a living. Employment creates opportunities that result in productive people contributing to the common good of society. Mining was and continues to be an important industry for all nations that maintain a vibrant economy.

Conclusion
The demise of Old Dinah is just one of many examples of the difficulties faced in transporting ore, equipment, and goods in remote arid environments during those early years. Conducting studies at historic sites in the desert is a learning process that can reveal much more than expected to the field observer. Understanding all aspects of historic sites, including botanical observations and other biological discoveries, can lead to further scientific and historic research.

Note. This research project, locating Old Dinah’s abandonment site, turned out to be a great learning experience in desert exploration. Conducting historic research and the planning of field work involved with locating the site forced us to learn about the interesting history of Death Valley. I want all of those reading this article to apply the same research process so that you too can learn in detail about Death Valley and its past. If I provided details of the site’s GPS location, the learning process would not be the same for the desert explorer.

References


Photo Sources: Historic black & white photo, public domain from the Internet. Color photo by Stephen Mulqueen taken November 9, 2019.
Excavation of a mammoth (*Mammuthus* sp.) from groundwater discharge deposits in Amargosa Valley, Nevada

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**Abstract**—Here we report the invertebrate and vertebrate paleontology of a paleontological site on Bureau of Land Management-administered land in the Ash Meadows area of Amargosa Valley, Nevada. A partial mammoth skeleton (*Mammuthus* sp.) was excavated from what are interpreted to be late Pleistocene groundwater discharge deposits older than 21,237-20,774 years cal BP. Groundwater discharge deposits across the Mojave Desert region are important proxies for global Pleistocene climate fluctuations. Upon completion of the excavation, three plaster field jackets were constructed, containing cranial and postcranial elements, including partial tusks. The disarticulation of the recovered cranial bones and tusks may indicate post-mortem trampling by another large mammal prior to burial. Analysis of the sediments and invertebrate fossil assemblages together suggest that this locality was along the margins of a perennial, marl-producing marsh with seasonally dry edges and influence from a spring channel.

**Introduction**

The late Pleistocene vertebrate fauna of Amargosa Valley, Nevada is not documented in detail (i.e., Paces et al., 1996) when compared to the Tule Springs local fauna (Springer et al., 2011) described from the neighboring Las Vegas Valley. A partial mammoth skeleton (*Mammuthus* sp.) was excavated under permit from the Bureau of Land Management (BLM) [BLM Permit N-94350] between 2016 and 2018 from a single locality in Amargosa Valley, Nevada (Figure 1). Here we report the invertebrate and vertebrate paleontology of this site and characterize the sedimentology associated with this excavation for the purpose of interpreting the depositional environment and paleoenvironment.

**Physical setting**

The excavation site is located within Nye County, Nevada, on BLM-administered land near the northern boundary of Ash Meadows National Wildlife Refuge (Figure 1). Ash Meadows remains today the largest continuous desert

![Figure 1](image_url)
wetland in the Mojave; however, land use surrounding Ash Meadows varied throughout the 20th century, including farming, peat mining, and ranching in the 1960s and 70s. As a result of these activities, the modern landscape has been largely dissected by stream channels (Denny and Drewes, 1965). Some of these springs are no longer actively discharging due to the rate of groundwater withdrawal exceeding recharge. Prior to excavation, the in situ fossils were located within a modern phreatophyte flat (Table 1).

### Geologic setting

The excavation site lies near the northern margin of the Ash Meadows 15’ Quadrangle. The geology of this quadrangle, including a geologic map, was described by Denny and Drewes (1965), in which they reported finding no fossils within the Quaternary deposits. Eolian transport strongly influenced the Quaternary sedimentology of the area, creating dunes, depositing sediment on wet surfaces associated with groundwater springs, as well as removing sediment by deflation.

Mojave Desert wetlands, notably in southern Nevada, were highly sensitive to global climate events throughout the Pleistocene, prompting alternating periods of wetland expansion and contraction (Spaulding, 1990; Quade et al., 1995; Pigati et al., 2019; Springer et al., 2015, 2017, 2018). The sedimentology, stratigraphy, and chronology of groundwater discharge deposits are meticulously described from the Las Vegas Formation in the neighboring Las Vegas Valley (i.e., Haynes, 1965; Springer et al., 2011; Springer et al., 2015, 2017, 2018). Groundwater discharge deposits in Amargosa Valley are distinct from the Las Vegas Formation; however, groundwater discharge deposits throughout the Mojave Desert region also capture highly sensitive responses to climate fluctuations during the Pleistocene (Pigati et al., 2019). The rare-earth-element signature of Ash Meadows groundwater indicate that most of the groundwater within the Amargosa Valley region is derived from relatively deep, regional, groundwater flow, rather than from local recharge (Johannesson et al., 1997).

### Excavation and methods

The in situ fossils were originally discovered several years prior to the excavation by a local resident who conducted an unauthorized excavation prior to alerting the BLM Pahrump Field Office. Permitted excavation began in the fall of 2016, involving many student and community volunteers. Backfill from the unauthorized excavation was removed at the start of the permitted excavation process, and a 3x4 meter grid was established (Figure 2a). Hand tools were used during excavation, such as awls and wooden dowels, dental picks, and brushes. Paraloid B-72 consolidants in acetone were used to consolidate exposed bone. The bone elements were localized in the center of the grid, with the tuskS articulated together, projecting down into the ground (Figure 2b). Upon completion of excavation in the spring of 2018, the field crew constructed three plaster field jackets (Figure 2c–d) for transport to the Las Vegas Natural History Museum (LVNHM), the designated repository for this excavation (LVNHM Repository number: BLM 2019.004). The stratigraphy of the excavation site was recorded during excavation (Figure 3).

Lab preparation of the vertebrate elements was conducted using hand tools and paraloid B-72 consolidants in acetone. For paleoenvironmental
reconstruction, two invertebrate fossil assemblages were collected, including both in situ material exposed during excavation and screen-washed sediments recovered from the plaster field jackets during lab preparation (Figures 4–5). Sediments were wet sieved through a fine and coarse mesh, dried for 72 hours, and then microfossils were picked using a stereo microscope.

Sedimentological observations
Observations concerning the sedimentology within the mammoth excavation quarry are recorded in Figure 3. Starting from greatest depth to the ground surface, there is a blocky, well-cemented, resistant, carbonate layer approximately 10 cm thick, overlain by ~15–20 cm of poorly sorted sediments, consisting of gravel and “popcorn” carbonate nodules within a sandy clay matrix; the clasts are sub-angular to sub-rounded and polymictic. The overlying ~8 cm is composed of massive gray-green clay with small, oxidized root voids, underlying ~20–30 cm of gray-white silt with rare granule-size, sub-angular to sub-rounded clasts, and abundant gastropod, bivalve, and ostracode fossils. A medium-to-fine-grained, white sand lens level with the mammoth fossils at the ground surface contains abundant mollusk shells that pinches out laterally.

Vertebrate paleontology
Field Jacket A contains the proximal portions of two articulated tusks within the incisive bone (Figure 4a–b). Each tusk measures approximately 17 cm in diameter, or a ~53.4 cm circumference at the proximal end. It is difficult to distinguish between partial tusks of mastodon (Mammut spp.) and mammoth (Mammuthus spp.); however, no mastodon remains have yet been reported from southern Nevada from the Pleistocene. Unfortunately, cheekteeth were not recovered; these are diagnostic for proboscidean genera and species.

Lab preparation revealed intact cranial bones that were disarticulated from the tusks (Figure 4b). The following cranial elements were identified: partial frontal, right and left partial parietal, incisive and partial maxilla, left zygomatic arch, and partial temporal. The zygomatic arch, which was articulated with the temporal bone in situ, was vandalized in the field, but was recovered before the main field jackets were constructed. The frontal bone within this partial skull forms a high and domed shape in its articulation with the temporal bone (Figure 4a–b), which is characteristic of mammoth skulls, versus the low and rounded profile of mastodon skulls. This specimen is likely a Columbian mammoth (Mammuthus columbi), as they are the only Mammuthus species documented in southern Nevada. The mammoth skull is broken along the proximal end of the alveolar process, and the cranial bones are flipped over, rotated 180° relative to the tusks, which remain articulated together (Figure 4a–b). This may indicate that the skull was trampled by another large animal prior to burial. In addition to postmortem breakage, many of the bone elements recovered from this excavation have damage from modern root staining and etching, and modern plant roots were found within the jacketed bone elements.

Field Jacket B contains the partial right scapula and several cervical vertebrae, including relatively complete C1 and C2, and vertebral fragments (Figure 4c). The dorsal margin of the C1 vertebra (Figure 10b) is relatively flat, which is characteristic of Mammuthus, unlike the rounded, almost concave dorsal margins of the lamina found in Mammut (Figure 7) (Olsen, 1972). The close association and articulation of the other cervical vertebrae to the diagnostic C1 leads us to identify the remaining vertebrae to Mammuthus. It is not possible to determine what happened to the remainder of the skeleton, notably
were identified using their descriptions in Burch (1982), including *Pisidium* sp., *Pyrgulopsis* sp., *Planorbella* sp., *Valvata* sp., and *Promenetus umbilicatellus*. Both aquatic and terrestrial mollusks were identified from the assemblage collected *in situ* from the mollusk-rich sand lens, including *Pisidium* sp., *Tryonia* sp., *Pyrgulopsis* sp., and *Vertigo* sp. (Figure 6) based on their descriptions in Burch (1982).

**Age of the excavation site**

Aquatic mollusks are known to be potentially problematic for $^{14}$C dating due to the intake of $^{14}$C-deficient, 'old' carbon from the water column during shell growth, which yields an age that is older than the actual age of the shells (Brennan and Quade, 1997; Pigati et al., 2004). Fully aware of this issue but desiring an approximate age for this locality, we obtained an AMS date on a shell of an aquatic gastropod *Planorbella* sp. (Fig. 4d), which was very abundant within the sediments surrounding the mammoth fossils. The $^{14}$C age of this shell was determined to be $22,990 \pm 97$ RCYBP (Lab#: D-AMS 019910). That date was calibrated using the IntCal20 curve on OxCal 4.4 (Bronk Ramsey, 2009; 2021; Reimer et al., 2013; Heaton et al., 2020), which yielded a calibrated age of 27,551 to 27,144 years BP.

Pigati et al. (2004) determined that some taxa of terrestrial late Pleistocene gastropods yield reliable $^{14}$C ages, such as the genus *Pupilla* (Pigati et al., 2004). We did not encounter this genus within this assemblage, but snails belonging to the closely related genus *Vertigo* are present (Fig. 5c) that are within the same family (Pupillidae) and have very similar morphologies and ecological characteristics (Pilsbry, 1948). We accumulated

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**Invertebrate paleontology**

Mollusks (Gastropoda; Bivalvia) and freshwater ostracodes (Ostracoda) make excellent paleoenvironmental indicators; many genera of Pleistocene snails, clams, and ostracodes are extant and have well-documented habitat preferences. Invertebrate fossil assemblages along with sedimentological observations can help establish paleoenvironmental interpretations, such as various groundwater discharge regimes (Table 1).

The invertebrate assemblage from the fossiliferous silt included ostracodes, gastropods, and bivalves (Figure 5). The two identifiable species of ostracodes, *Candona cerca* and *C. averta*, were identified from their description in Forester et al. (2017) and are surface-dwelling wetland species. Fully aquatic mollusks from this assemblage the mandible, cheek teeth, and other postcranial elements. These elements may have weathered away or been transported prior to burial.

Field Jacket C was a small, ~30cm x 15cm jacket containing several unidentifiable bone fragments (Figure 2a).

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**Figure 3.** Stratigraphic column and interpretations of groundwater discharge regimes of the excavation site, showing the lithology of the excavation quarry.

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**Figure 4.** a) Diagram of a mammoth skull; shaded portions represent elements that were recovered and identified throughout excavation and preparation. Photographs of Jackets A (b) and B (c) after lab preparation. The tusks in alveoli (b) are rotated 180° relative to life position, so that the distal ends of the tusks are pointing toward the skull.
33 mg of Vertigo shells for $^{14}$C dating, which yielded a $^{14}$C age of 17,394 ± 57 RCYBP (Lab#: D-AMS 045565). This date was calibrated using the IntCal20 curve on OxCal 4.4 (Bronk Ramsey, 2009; 2021; Reimer et al., 2013; Heaton et al., 2020), which yielded a calibrated age of 21,191 to 20,830 years BP. Since these fossils were found in a sand lens that was deposited after the vertebrate fossils accumulated, the mammoth fossils are likely older than 21,237 to 20,774 years cal BP, at the end of the Last Glacial Maximum.

Discussion and interpretation of depositional environment

The sedimentology and stratigraphy of this site reflect a change in wetland environments through the stratigraphic section (Figure 3). We interpret the resistant carbonate layer stratigraphically below the mammoth fossils to be water-table carbonate (Table 1) that formed through evapotranspiration within the wetland sediments. The poorly sorted sandy clay with angular-subangular gravel and carbonate nodules suggests the presence of an ephemeral wash or spring channel that deposited larger clasts and weathered carbonate nodules that formed from the underlying carbonate layer. It is common to see reworked carbonate rubble within stream deposits in the Las Vegas Formation (Springer et al., 2018). The green clay with root voids and the gray-white blocky carbonate-rich silt with abundant mollusks and ostracodes are indicative of wetland marl facies (Table 1). The invertebrate assemblages from this locality suggest that it was located along the periphery of a perennial marsh that had input from an ephemeral wash or stream channel. The two identified species of ostracodes (Figure 5) were originally described from pale green calcareous mudstone with interbedded channel sands with abundant mollusks at Corn Creek, Nevada (Quade, 1986; Forester et al., 2017). These sediments at Corn Creek are interpreted to have been deposited in perennial, well-oxygenated wetlands with seasonally dry margins (Quade, 1986), which is consistent with the interpretations of the sediments at this excavation site. The transition in depositional environments of the Las Vegas Formation following the end of the Last Glacial Maximum after ~23,000 years BP reflects a shift from extensive wetlands to point-source discharge and ponding (Springer et al., 2018).

Interpretations cannot be made about the disturbed sediments above the silty clay with mollusk shells due to reworked backfill from the unauthorized excavation conducted several years prior.

Groundwater discharge deposits and lacustrine deposits can have similar lithologies, so it is important to clarify how the distinction was made at this locality. Pleistocene lacustrine deposits are found in the Mojave Desert (i.e., Kirby et al., 2015); however, they are...
typically confined within a closed basin, or are part of a river drainage system (Pigati et al., 2014). This locality is not within a closed basin, but rather is on the floor of the valley, which drains into the Amargosa River. Additionally, the depth to groundwater table is shallow (<2 m) and there are several actively discharging springs and phreatophyte flats located close to the excavation site. The fossil invertebrate assemblages identified from this site are taxa associated with spring discharge and not lacustrine environments (Figures 4-5). The tectonic setting of this locality is also inconsistent with a lacustrine interpretation; the expression of shallow faults along the Stateline Fault System and linear spring alignments (Figure 1) strongly support these interpretations (Guest et al., 2007).

Summary and conclusions
This locality was studied to document the excavation of a partial proboscidean (cf. *Mammuthus columbi*) skeleton from 2016 to 2018 in Amargosa Valley, Nevada and to record the sedimentology and vertebrate and invertebrate paleontology to interpret palaeoenvironmental conditions. Excavation and lab preparation yielded three plaster field jackets containing both cranial and post-cranial skeletal elements. We recovered the partial frontal, right and left parietal, incisive bone, and partial maxilla, left zygomatic arch, and partial temporal bone as well as largely complete scapula and several cervical vertebrae (Figure 4). The disarticulation of the cranial bones and tusks may indicate post-mortem trampling by another large mammal prior to burial, causing the skull to fracture and rotate along the nasal opening so that the anterior end of the skull faces the distal end of the tusks. The cranial and post-cranial elements also show modern root staining and etching, as well as Stage 1-2 surface weathering, which suggests they were exposed for less than six years prior to burial (Behrensmeyer, 1978).

The sedimentology and stratigraphy of this locality suggest a shift between stream facies and marl-producing wetland facies (Figure 3) (Table 1). The sediments and invertebrate assemblages together suggest that this locality was along the margins of a perennial, marl-producing marsh with seasonally dry edges and seasonal influence from spring channels or ephemeral washes. This marsh would have been fed by slowly flowing groundwater with localized ponding. The *in situ* terrestrial gastropods collected from the sand lens cross-cutting the vertebrate fossils dated to 21,237-20,774 years cal BP, suggesting that the vertebrate fossils were deposited before this time. Although the fossiliferous groundwater discharge deposits of Amargosa Valley require more formal and comprehensive description and analysis, it is still important to document individual paleontological sites to incorporate them into a regional context.

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How far can the scoria go?: transport of scoria from the cinder cones to Soda Lake basin in Mojave National Preserve

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Scoria originating from the field of cinder cones in Mojave National Preserve can be observed over 10 miles away from the cones as well-rounded, pebble- to cobble-sized deposits on top of sandy flats near the low-elevation ephemeral Soda Lake. How was the scoria transported to these locations? Under what conditions? We investigate the mechanics of transport of scoria in an arid environment by focusing on the following research goals: 1) interpret method of transport and 2) estimate magnitude of flow. Our methodology includes a combination of field observations and modeling. First, we interpret that the scoria was fluvially transported based on sedimentological evidence, including the shape of the deposited lobes from sheetfloods and the well-rounded shape of the scoria cobbles. Since the cobbles are heavier than water, the scoria must have been transported within the water body of a sheetflood and crepted and/or saltated across the surface. This transportation process rounded the scoria. Second, we estimate the magnitude of flow required to mobilize scoria cobbles; this goal was completed by measuring each rock mass, coefficient of static friction, and submerged weight of 30 scoria cobbles. We hypothesize these scoria-carrying floods formed from monsoon-like, short duration, intense summer rain events that we modeled using KINEROS2. Modeling results parameterized with digital datasets for topography, soils, and land cover available via the Automated Geospatial Watershed Assessment tool indicate that rainfall events sufficient to mobilize scoria cobbles are rare and intense.

Copper Mountain College, an unassuming host of ancient wetlands and deformation along the Pinto Mountain Zone

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Copper Mountain College is located in a geologically diverse and tectonically active area near the junction of the Mojave Desert, Eastern Transverse Ranges, and San Bernardino Mountains tectonic blocks. The campus lies along the east-striking Pinto Mountain Fault (PMF), a regionally important left-lateral fault zone that is intersected from the north by right-lateral faults of the Eastern California Shear Zone. Although not well exposed in the youngest alluvial deposits, just NW of Copper Mountain College, the fault forms a scarp in deposits dated by Infrared Stimulated Luminescence (IRSL) at 1450 ± 50 yr. Directly north of the PMF is Copper Mountain (CM), which is a faulted and sheared bedrock block composed of Precambrian (?) gneiss, Triassic intrusive rocks, and Jurassic diorite and quartz monzonite. The block is bounded by the NW-striking right-lateral Copper Mountain and Calico-Hidalgo faults to the west and east, respectively. The campus is surrounded by Quaternary surficial deposits typical of desert environs. To the south, north-prograding alluvial-fan deposits advance from drainages in Joshua Tree National Park. Playa sediments occur in lowlands west and northeast of CM. Eolian sediments are locally abundant, and form sheets, ramps, and dunes. The oldest Quaternary deposits in the Copper Mountain College area crop out in an ~30-m-high structural pop-up just NE of campus. This pop-up formed either within a right step in the left-lateral PMF, or at the intersection between the PMF and CMF. Sediments exposed within the pop-up dip as much as 85° southward. They typically are fine-grained fluvial sediments with sparse cobble-boulder horizons; clasts consist of plutonic rocks derived from the south, rocks local to CM, and quartzite and Pioneertown basalt derived from the west. Near the top of the pop-up, sediment low in the section is dated by IRSL at 106,000 ± 9820 yr. Newly recognized in this study are Quaternary groundwater discharge (GWD) deposits that partly underlie Copper Mountain College. The GWD deposits consist of two sections of massive, very fine sand and mud that fine up to clayey mud and are capped by hard carbonate. IRSL dates obtained on GWD deposits are as young as ~10 ka and as old as 97,290 ± 8200 years.

Corals and volcanics: revisiting a forbidden zone—Barrett Canyon, in the Carrizo Impact Area of the Anza-Borrego Desert State Park®, California

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Some of the most scenic and intriguing places in the Anza-Borrego Desert State Park® are in the Carrizo Impact Area (CIA) – closed to the public since its use as a bombing range beginning in 1942. Few researchers have obtained the permits necessary to visit this area. The scenic areas that are closed include Red Rock Canyon,
Lavender Wash, Deguynos Canyon and Barrett Canyon. Local photographer Ernie Cowan was granted access to the CIA in the mid-2000s, providing us with a recent glimpse of its formations and the fossils they contain. Barrett Canyon sediments are assigned to the Imperial Group.

A small collection of fossil corals and mollusks was sent to Prof. T. Wayland Vaughan in the fall of 1903. The materials were so exceptional that a paper was presented at the Geological Society of Washington in January 1904 and an expedition was launched later that month.

What made these materials so exceptional was what they represented—a shallow tropical warm sea was the predecessor to the modern cold water Gulf of California environment. The corals in the collection were not known to exist along the Pacific coast but did exist in the Caribbean.

During that expedition, corals were collected from Barrett Canyon, Alverson Canyon, and other areas in the Coyote Mountains. Fourteen specimens of Barrett Canyon coral are in the US National Museum, Smithsonian Collections, with eight designated as species holotypes.

The Barrett Canyon corals lie unconformably atop a large, 62 m-thick (200 ft) lava flow. Effusive lavas, pillow basalts, cinder cones and lahars are present throughout the area. These rocks are evidence of subduction-related volcanism and reflect the transition from subduction to rifting which opened and created the tropical Imperial Sea. Nearby Alverson Canyon andesites and olivine basalts are dated at 17 Ma.

The coral reef at the head of Barrett Canyon is estimated to be 500 m long, 20 m wide and 1.2–2.0 m high. It formed during the Miocene, creating quiet lagoons behind the sea-facing reef front. The faunal list for Barrett Canyon is a who’s-who of tropical marine mollusks, gastropods, echinoderms, and corals. There are five families of corals, two dozen families of gastropods, two dozen families of bivalves, and a half dozen species of echinoderms, bryozoans, barnacles, and crabs.

In addition to the US National Museum, Barrett Canyon fossils are housed in the collections at Anza-Borrego Desert State Park® Stout Research Center, San Diego Natural History Museum, State University of Iowa, the University of California Museum of Paleontology, and the California Academy of Science. Marine fossils from Alverson Canyon and other areas in the Coyote Mountains are also present in those collections.

Tropical marine life disappeared with the uplift of Central America that ended about 3 Ma. Warm Caribbean waters were replaced with the cold Pacific, the tropical climate became temperate, and water circulation patterns changed. The area transitioned to a terrestrial environment with the arrival of the Colorado River delta sediments.

Very little is known about the Barrett Canyon volcanics and their relationship to flows found nearby in the Coyote Mountains and Volcanic Hills, the Jacumba volcanics to the west and the Colonia Progresso volcanics to the south. As USGS geologist Walter Mendenhall said in 1910, the Coyote Mountains “exposures and structures are so clear that its geology will be an open book to the fortunate student to whom falls the pleasant task of deciphering it in detail”.

A framework for mapping volcanic cones, ash and basaltic flows in the Anza Borrego Desert State Park®
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Individual volcanic features including ashes, basaltic lava deposits and cones have been described in the context of geologic and paleontologic dating within the Anza Borrego Desert State Park (ABDSP). The early to late Miocene in ABDSP is typified by volcanism, the formation of fault bounded rift basins, and high-energy terrestrial deposition (Red Rock Formation, Alverson Canyon Formation, and Split Mountain Group). We have created a framework which allows study and mapping of these features.

The starting point for this effort has been to gather previous references and maps from scientific publications and from the ABDSP’s own databases. The ABDSP paleontology database contains 29 references to volcanic ash. These are in Coyote Badlands (Ash Wash), Borrego Badlands, and Arroyo Tapiado.

Dozens of volcanic features are identified in Keyhole Markup Language (KML) format files. These cover areas such as Borrego Palm Canyon 7.5' Quad and Collins Valley 7.5' Quad, while others are specific to particular field surveys. The KML files are readily overlaid on digital maps, including Google Earth and ArcGIS maps maintained by the ABDSP.

A comprehensive bibliography of papers and theses published over the past fifty years has been compiled and maintained up-to-date. This includes individual Portable Data Format (PDF) files for many papers. The studies include descriptions of Alverson Formation, Tapiado Ashes, Viejas Formation and the Bishop Tuff.

These resources are available to staff, volunteers and visiting scientists at the ABDSP. Each of the elements in the framework are being updated and revised as required to insure completeness and accuracy.

Comparing observed vs estimated mammalian diversity in the middle Miocene Dove Spring Formation
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The complex paleotopography and excellent fossil record of the Mojave region make it an ideal location for testing hypotheses about the influence of the landscape on mammals. By placing a well-resolved biochronologic record against the timing of major tectonic or climatic
intervals, it is possible to evaluate relationships between landscape development and changes in faunal composition. Here we present a revised biochronology of southern California's Dove Spring Formation based on current magnetostratigraphy.

We studied the fossil record in the Dove Spring Formation (12.5 - 8.0 Ma) from the El Paso Basin to generate stratigraphic and temporal ranges for over 100 species of large mammals within 0.5-Myr time intervals. We determined 80% confidence intervals for the temporal range of each species to evaluate the uncertainty of their stratigraphic ranges, as the true residence time of many species is underrepresented due to the incomplete nature of the fossil record. For each 0.5-Myr interval, we calculated the standing diversity and per-capita rates of origination and extinction (represented here by arrivals and disappearances from the basin's fossil record). For the fossil community as a whole, a peak in originations occurs 10.5 Ma, followed by a peak in extinctions at 9.5 Ma. A peak in estimated mean standing diversity at 10.0 Ma coincides with the appearance of numerous rare taxa, suggesting a preservational bias. However, the 11.0 to 10.5 Ma peak per-capita origination rate for large mammals precedes a major increase in fossil productivity, indicating the arrival of numerous species to the basin at that time. Per-capita extinction rate peaks at 9.5 Ma, which coincides with the highest locality frequency and number of specimens.

These changes in diversity may have been influenced by tectonic episodes. Beginning near 10.0 Ma, the El Paso Basin was rotated and translated westward along the Garlock fault, followed by extension and increased subsidence initiating near 9.0 Ma. A sustained low sediment accumulation rate is associated with this interval, suggesting an increase in basin area or a change in relief from the source area. Barriers to dispersal were likely removed and habitat availability increased for ungulates. Based on our current data, these tectonic changes had no apparent influence on preservation conditions in the basin, as the majority of the section consists of taphonomically similar sedimentary facies. Further investigation of depositional environments will provide independent estimates of preservation rate and allow us to distinguish high taxonomic rates from preservational artifacts.

**Titus Canyon Formation, Death Valley National Park, California and Nevada—Tuffaceous marker beds, \(^{40}\text{Ar}/^{39}\text{Ar} \) dates, and constraints on age of Titus Canyon Fauna**

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The continental Titus Canyon Formation (Fm) is exposed in the Grapevine Mountains and the adjacent northwestern third of the Funeral Mountains to the southeast in northeastern Death Valley National Park, southeastern California and contiguous southwestern Nevada. It consists of a number of successive lithostratigraphic units, including (1) a lower monolithologic limestone breccia unit, (2) an upper heterolithologic limestone breccia unit, (3) a quartzite conglomerate unit, (4) the fossil land mammal-bearing lower red beds, (5) the first green conglomerate unit, (6) the variegated member (mb), (7) the middle sandstone and conglomerate beds, and (8) the upper “red beds.”

Four tuffaceous marker beds, all visible in Google Earth satellite imagery, have been especially useful in recognizing unit boundaries and in bracketing the fossil-bearing interval of the lower red beds from the western fork of Titus Canyon to upper Titanothere Canyon in the southeastern half of the Grapevine Mountains. They include an unnamed tuffaceous sandstone layer and the successively overlying California Institute of Technology (CIT) locality (loc) 255, Monarch Canyon, and Unit 38 Tuff Beds. The first or lowest tuffaceous layer, a comparatively thin bed of light yellowish-gray tuffaceous sandstone, lies at the base of the lower red beds in the southwestern reentrant of Titus Canyon’s western fork. There, the layer (1) immediately overlies an interval of sandstone beds and conglomerate lenses that occurs locally as the upper part of the quartzite conglomerate unit, and (2) directly underlies a thicker, dark reddish-brown mudstone layer representing the lowermost red bed and fossil-bearing layer of the lower red beds in the western fork. One of three quarry sites constituting CIT loc 254 (Quarry 1 herein) occurs in that basal red bed. The corresponding assemblage represents one of the three oldest in the Titus Canyon Fauna. Quarries 2–3 are situated in a much thicker, bright reddish-orange fossil-bearing mudstone layer lying two red beds higher in the local succession.

In the northeastern portion of Titus Canyon’s western fork and in upper Titanothere Canyon, the second or CIT Loc 255 Tuff Bed occurs between the first and second red beds below the top of the lower red beds. The CIT loc 255 quarry site and University of California Museum of Paleontology loc V87019 lie in the second red bed and yielded the youngest assemblages of the Titus Canyon Fauna. Consequently, the fauna is bracketed by tuffaceous marker beds. Based on the shared occurrences of 12 extinct land mammal species, the Titus Canyon Fauna is a correlative of the late early late Duchesnean (= late early Du2) Upper Porvenir Local Fauna (LF). The latter assemblage occurs 0–26.8 m (0–88 ft) above
the lower marker bed (i.e., in Blue Cliff Horizon) in the lower third (but not lowermost part) of the Chambers Tuff Fm in Trans-Pecos Texas. Critical age-diagnostic taxonomic occurrences include (1) regional last appearance datums for the ischyromyid Quadratominus? gigas, the amphibycnophi Daphoenticis n. sp. (small), the brontotheriid Protitanops curyi, the helaeodont Colodon stovalli, the rhinocerodont Telestaceras mortivallius, perhaps the agriochoerid Eomynex transmontanus, and the leptomyercyd Hidrostereum transpecosensis, and (2) the regional first appearance datum of the cylindroodont Dolocylindrodon texanus. The Chambers Tuff Fm is bracketed by the Buckshot Ignimbrite and Bracks Rhyolite. The two volcanic units were determined to be 37.80 ± 0.15 Ma and 36.67 ± 0.08 Ma, respectively, based on 40Ar/39Ar single-crystal laser fusion (SCLF) analyses of sanidine crystals and the previously accepted date of 520.4 Ma for the McLure Mountain hornblende (MHHbl-1) flux monitor. The age of the Buckshot Ignimbrite was subsequently determined to be 37.68 ± 0.04 Ma by Christopher D. Henry (Nevada Bureau of Mines and Geology) using the currently acknowledged age of 28.201 Ma for the Fish Canyon Tuff sanidine (FCTs) flux monitor. The date for the Bracks Rhyolite is recalculated herein using ArArReCalc as 37.15 ± 0.08 Ma to ensure compliance with the same calibration standard. When compared to dates for subepochal boundaries in the Geologic Time Scale 2020, those dates indicate (1) that the Chambers Tuff Fm, the Upper Porvenir LF, and, by correlation, the lower red beds are at least somewhat older. It was underlying the lower red beds are at least somewhat older. Fauna are earliest late Eocene in age, and (2) that the units by correlation, the lower red beds and the Titus Canyon Fm is bracketed by the Buckshot Ignimbrite and the presently recognized date originally determined to be 34.3 ± 0.7 Ma in age, based on the northwestern end of the Funeral Mountains. It was underlying the lower red beds are at least somewhat older. Fauna are earliest late Eocene in age, and (2) that the units by correlation, the lower red beds and the Titus Canyon Fm is no younger than earliest late Eocene or late early Du2 in age, whereas the formation is early Oligocene or Wh2 at its top.

Field efforts in the Grapevine Mountains were conducted under United States Department of the Interior, National Park Service Scientific Research and Collecting Permit No. DEVA-2017-SCI-0037. It was issued to Nyborg, Lander, and Dr. Kevin E. Nick (Loma Linda University).

Variable factors affecting camera trapping for two ground squirrel species in the western Mojave Desert

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Between 2017 and 2021, the author monitored motion surveillance cameras that were installed to census the occurrence of the State-listed Mohave ground squirrel (Xeroperomys philomohavensis) (MGS) at three sites in the western Mojave Desert. MGS, as well as white-tailed antelope ground squirrel (Ammonoperomys leucurus) (AGS), are the focal squirrel species of this study. Numerous factors affect the function of motion cameras and the subsequent review and analyses of photographs. Variable factors considered in this presentation include: 1. hardware, 2. bait presentation, 3. animal identifications, 4. weather conditions, and 5. collecting/reporting results. Camera quality may affect results, but even two cameras of the same make vary. Bait presentations include blocks, PVC tubes, and screened sandwich containers, each with pros and cons. Between the two squirrel species within the study area, there are difficulties discerning both among individuals and between species. Early indications suggest that winter precipitation, which affects timing of annual plant germination, likely affects detectability and abundance of squirrel images. As camera trapping for these species is relatively new, since 2011, variable approaches are being implemented until acceptable methods are identified.
The future of the moving mud spring, Imperial County, California

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This report follows on several past papers about the moving spring. The spring moved about 100 meters in three years, overrunning critical infrastructure that required rapid response and expensive mitigation by Union Pacific Railroad, Kinder-Morgan, AT&T and CalTrans. Based on changes in water production and sediment load, we have modeled the spring’s behavior using simple physical processes that may predict its future. It appears that the spring’s activity has been diminishing and will continue to diminish and the spring will disappear or become little more than a trickle by 2024–2028. There is, however, evidence that the spring is part of a larger, slowly changing regional aquifer that may yet produce unexpected emergence of surface water in the area.

Time-space patterns of latest Quaternary surface ruptures on the eastern Pinto Mountain Fault near the intersection with the southern Mesquite Lake Fault, Twentynine Palms, California

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The Pinto Mountain Fault Zone (PMFZ) marks a major structural boundary between two structural domains: east-oriented sinistral faults of the eastern Transverse Ranges (to the south) and northwest-oriented dextral faults of the Eastern California Shear Zone (ECSZ) (to the north). These sinistral and dextral domains intersect in the Twentynine Palms area where the U.S. Geological Survey (USGS) is conducting detailed geologic mapping and geochronologic investigations designed to clarify geometric, kinematic, and temporal interactions among the two domains. The USGS has focused much of this research specifically on the eastern section of the left-lateral PMFZ near the intersections with the right-lateral Mesquite Lake Fault Zone (MLFZ).

Methods—Detailed geologic mapping in the Twentynine Palms area provides a foundation for accompanying geomorphic and geochronologic analyses of these faults. We mapped using a combination of field traverses and photointerpretation to recognize and map faults and Quaternary surficial units and to determine which map units are or are not displaced by faulting. Faults are most easily identified in road and stream cuts, where they are expressed as discrete fracture-shear surfaces or zones. We extended the faults across the landscape using a variety of criteria, including low fault scarps and larger-scale fault-bounded escarpments, structurally dismembered and (or) bounded ridges, laterally offset drainages, fault exposures in transversely oriented incised drainages, and structural alignments of vegetation and composite topographic features. Samples for luminescence dating were obtained where mapping identified the need for stratigraphic control on map units and structural control for faults and folds. Morphometric landscape analysis integrates these two data sets.

Pinto Mountain Fault Zone—From its intersection with the MLFZ, the PMFZ extends ~60 km westward to its intersection with the San Andreas Fault. Along this reach, evidence from basement rocks suggests cumulative left-lateral displacement on the order of 16–20 km, but the precise timing and possible periodicity of sinistral-slip events has not been determined. Our mapping clearly demonstrates significant mid- to late Quaternary slip activity on the 60 km western and central reaches of the PMFZ.

Near the western margin of Twentynine Palms at the eastern edge of this study area, the surface trace of the PMFZ is characterized mainly by a generally narrow, structurally simple system of one to several subparallel to overlapping individual fault strands. However, eastward along a 10-km distance within southern Twentynine Palms municipality, the PMFZ increases both in width and complexity and the zone also begins to curve (bend) stepwise southeastward as it approaches the intersection with the MLFZ. At the west end of the study area the fault zone has a total width of surface rupture <100 m, which progressively increases to >5 km near its intersection with the MLFZ, thereby forming what we refer to as a composite fault zone. This eastward-widening of the composite fault zone is achieved by stepwise additions of fault strands along its south margin as older strands to the north are abandoned. New strands commonly initiate at en-echelon to branching right steps in the composite fault zone; the new strands have a more southeasterly orientation than do their counterparts west of the right steps. The composite fault zone consists of discrete internal zones of complex faulting and deformation that progressively increase from one to 4–5 in number and from <5 m- to >100 m in width eastward across the map area.

Within the composite fault zone, individual faults bound blocks with little internal deformation. Each of the internal fault zones comprises in detail a complex array of subparallel to branching shear and gouge zones, 2–8+ in number and centimeters to meters in width, that collectively represent zones of distributed surface rupture. Many internal fault strands in the composite zone display geomorphic and structural features (vertical scarps, fault-bounded escarpments and popup structures, oblique slickenlines, net displacements, and...
folded or tilted strata) that indicate significant secondary transpression, commonly localized at SE-striking, rightshifted, constraining bends, branches, or step-overs. The secondary deformation increases eastward both in intensity and in aerial coverage as the PMFZ approaches the MLFZ.

Mesquite Lake Fault Zone—The NW-striking MLFZ extends for at least 40 km from the Twentynine Palms Marine Corps Air/Ground Combat Center southeastward through Mesquite Lake playa to the fault’s intersection with the PMFZ adjacent to the northern mountain front of the eastern Transverse Ranges. The fault forms one to several conspicuous traces across youthful Quaternary deposits of Mesquite Lake playa, traverses the east margin of an extensive eolian-dune field, and crosses alluvial fans at or near the base of a prominent fault-generated uplift. Locally we have observed individual geologic and geomorphic features (e.g., fluvial bars and channels on alluvial fans) that are dextrally displaced as much as 3–4 m; we interpret these as evidence for the most recent ground-rupturing earthquake event (i.e., most recent event, MRE) on this fault.

Late Quaternary faulting history—We are beginning to work out the timing and distribution of late Quaternary displacements on the PMFZ and MLFZ, as well as how interacting sinistral and dextral slip within these two zones has influenced the geometry and coeval structural evolution of both. To support this analysis, we acquired 18 new luminescence dates by sampling late Quaternary surficial materials disrupted by, or concealing, various mapped strands of the two zones. Our working model for fault-movement history includes the following elements:

1. Paleoearthquake succession: Using offset surficial map units as well as displaced individual geomorphic features, we have been able to demonstrate that multiple surface ruptures related to paleoearthquakes have occurred on various fault strands over the last 60 ka or so. Pinto Mountain Fault Zone: We find that Holocene surface-rupturing paleoearthquakes have preferentially occurred on two to three southern and southeastern strands of the PMFZ, in contrast to only late Pleistocene ages for youngest rupture events on the more northern PMFZ strands. We have particularly close constraints on the latest Quaternary rupture events at 3 sites on the southern branch of the PMFZ that displace a stacked series of 4 alluvial deposits which collectively establish a latest Holocene age for the most recent surface rupture. Exposed fault-stratigraphic relations include: (a) oldest deposits with estimated mid-late Pleistocene ages extensively sheared by multiple (many more than 3) faulting events: (b) an intervening deposit with a latest-Pleistocene age of 12.3 ka displaced by ≥ 2-3 events; (c) several overlying upper deposits with 1.4 ka to 5.1 ka ages displaced by the single youngest event; and (d) an overlying uppermost unfaulted deposit with an age of 1.3 ka. Farther to the southeast, three late Holocene alluvial fan deposits with ages between ~1.4 ka and ~3.1 ka display a series of remnant depositional bars and swales that are sinistrally offset 3-4 m by a single event on one strand of the PMFZ, probably related to the MRE at the sites on the fault zone to the west. Many late Pleistocene alluvial fans with numeric ages between ~15.0 ka and ~49.6 ka are highly deformed and laterally displaced ≥ 20 m by multiple events on all branches of the PMFZ.

Mesquite Lake Fault Zone: We have evidence for multiple late Quaternary surface ruptures on the MLFZ, with at least one event well constrained to a latest Holocene age. Sheared eolian-dune sand with a late Holocene date (~1.3 ka) occurs at the base of a faulted margin of a dunefield sag playa in the northern Twentynine Palms area. To the SE along this fault, several nested alluvial fans with ages of ~2.0 ka and ~2.3 ka are right laterally offset by 3–4 m. These zones of surface rupture are likely related to and define the MRE on this part of the MLFZ, which is very similar in timing and displacement size to the MRE defined for the PMFZ. Many earlier surface ruptures are indicated on the MLFZ by a structurally isolated, right-laterally displacement > 50 m of a late Pleistocene alluvial fan with ~31.0 ka dates.

2. Interaction between PMFZ and MLFZ: Collectively these numeric ages for the two intersecting fault zones indicate very similar timing and displacement characteristics for at least the youngest paleoseismic surface ruptures on these two nearly conjugately oriented structures. This suggests the occurrence of temporally clustered to possibly coseismic MREs on both major faults which resembles the paired surface ruptures on two intersecting orthogonal faults in the 2019 Ridgecrest earthquake sequence in the ECSZ to the north.

Applying GIS to a fossil collection: past, present, and future

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Distribution of fossil taxa through both time and space is difficult to demonstrate in a visual format. Traditionally, fossils are plotted within a stratigraphic column or on a bar graph that is usually delimited by increments of time or distance. GIS technology allows precise placement of fossil localities within a geographic area. Adding geologic formations provides representation of the depositional environments.

For the Fish Creek-Vallecito Creek section, plotting coordinates for core samples allows display of the geomagnetic polarity sequence for its 5 km-thick section. Dated ash beds serve as markers for calibrating the sequence to the global polarity time scale. The unique physical context of this section in Anza-Borrego Desert State Park makes it possible to bring all these data together into a single mappable framework that allows dating sources, geologic formations, geographic location, and stratigraphic position of individual fossils to be
displayed atop a photographic satellite image. This format allows for: analysis of the fossils already in the collection (first and last appearance, macro- and micro-habitats), validation of catalog data (anomalous occurrences, associated fossils in neighboring sites), and planning field activities (outline unexamined areas, focus searches based on reasonable expectation of discovery of target taxa, identify areas needing localized higher resolution study such as paleomagnetic surveys or detailed geologic mapping).

**Late Pleistocene invertebrate fauna, Lake Elsinore area, Riverside County, California**

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Over the past dozen years, late Pleistocene freshwater fossil mollusks have been discovered in two separate construction projects along the western end of the San Jacinto River in Riverside County. Each of these projects, less than five miles from one another, was located next to a lake: Canyon Lake and Lake Elsinore. Both sites had in common a few species of freshwater snails that still live in southern California today, *Physella* sp. cf. *P. heterostropha*, *Planorbellata tenuis*, and *Gyraulus parvus*, as well as a native land snail that lives in very moist areas (often at the edge of a body of water), *Succinea rehderi*. Only two shells of *Succinea* were found at Canyon Lake, but hundreds of specimens were collected at Lake Elsinore. The native freshwater mussel, *Anodonta californiensis*, only occurred at Lake Elsinore. Three species of freshwater gastropods that today are most common in Canada and the northern United States, *Stagnicola elodes*, *Gyraulus* sp. cf. *G. hornensis*, and *Valvata humeralis*, occurred at Canyon Lake with just two very small specimens of *Valvata humeralis* at Lake Elsinore. The calibrated age for the Lake Elsinore site is 18.1 cal ka BP, and for Canyon Lake is 13.0 cal ka BP.

**An early-historic-age bison and pronghorn skinning and butchering site in the Great Basin Desert of northern Nevada**

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We report the recovery and preliminary analysis of an assemblage of Bison bison bones, along with a smaller number of pronghorn (*Antilocapra americana*) bones, from a recently discovered fossil site in the Great Basin Desert. The site is on private land a few km east of Carson City, Nevada. Several of the bones exhibit distinct cut marks, some of which appear to have been made with a steel tool. The excavation site, which we have named the Gordon Bison Butchering Site, lies within the floodplain of Clear Creek, a tributary of the Carson River. There is no previous record of Bison of any age in the Carson Valley, so this occurrence extends the range of Bison into this region.

We have recovered bones representing at least four individual bison and one pronghorn from multiple excavation pits up to about 50 m apart. A subadult bison from Pit #1 was almost completely articulated, while the others are represented by isolated bone elements. All of the bones were found no deeper than approximately 80 cm below the ground surface. The Pit #1 bison was lying in a peculiar orientation, with its ventral side down. Its left foreleg was projecting forward from the shoulder, and its right foreleg was projecting downward into the floodplain sediment. Hundreds of pupal casings (puparia) of the black blow fly (*Phormia regina*) were found in the sediment around the rib cage. We infer that this animal was skinned and butchered, but not dismembered. When skinning a bison, Native Americans would typically orient the animal on its stomach and slit the skin down the midline of the back, in part to recover the highly desirable hump meat. The orientation of the Pit #1 skeleton is compatible with this interpretation. The tail was often removed with the skin, which may explain the absence of caudal vertebrae in this specimen. The abundant black blow fly puparia record a relatively long (several days) exposure of the skinned, butchered carcass to flies. However, the lack of dismemberment of this skeleton indicates that it was buried relatively quickly, before carnivores were able to find it.

Radiocarbon dating indicates two possible age windows for these fossils: (1) around the year 1700, or (2) sometime during the nineteenth or early twentieth centuries. The inferred use of steel tools eliminates the earlier age possibility. Therefore, this site probably dates to the mid-to-late nineteenth century, by which time European Americans had introduced steel tools into the Carson Valley region.

Our preliminary interpretation is that this site was a Washoe hunting camp that was active in the spring, prior to peak snowmelt in the nearby Sierra Nevada. Spring runoff then caused Clear Creek to flood, burying the site in floodplain sediment and sequestering the bones from carnivores. Our analysis of high pupal mortality recorded in the black blow fly puparia is compatible with this scenario.

**First Pleistocene record of *Equisetum* from the Mojave River**

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Swinerton Renewable Energy has been constructing a photovoltaic solar project on the west side of the Mojave River near Victorville north of the Southern California
Logistics Airport. This is on the eastern edge of the George Surface, as preserved in that region. The George Surface is a late Pleistocene river terrace representing the broad floodplain of the ancestral Mojave River. This name derives from the former George Air Force Base located on that feature and the name was proposed in 1999 by Cox and Tinsley. Borings for a gen-tie power line (connecting the solar field to an existing substations) were monitored in February of 2021. All of these were in Pleistocene terrace deposits of the Mojave River. The upper fluvial unit of the Mojave River began aggrading between 475 and 575 ka and ended 60 or 70 Ka, according to magnetostratigraphy and OSL. The Mojave River then began down-cutting in its present valley. Exposures of perhaps 120 feet (= 36 meters) of Pleistocene sediments can be seen on the west side of the river near the solar project. This particular gen-tie structure site is at 2,657 feet asml. The elevation of the adjacent George Surface is approximately 2,800 feet. Borings have revealed 185 meters of Plio-Pleistocene sediments in the Mojave River Valley in the Victorville area. The upper 110 meters are fluvial sediments.

An auger with a diameter of 6 feet brought up remains of Equisetum sp. in iron oxide cemented sands 20 feet bgs. This is relatively near to the top of the sequence. One specimen consists of two longitudinal halves of a stem preserved in horizontal position and measures 4 mm at its maximum width. It shows definite septate nodes. Another stem is preserved in vertical position in the same lump and has a maximum diameter of 7 mm. A third is preserved in another lump and seems to transgress a layer of fine sand and another of coarse sand. Its maximum diameter is 4 mm. Given the dates of the aggrading sequence, the fossils must be no younger than 60 Ka. There has been significant erosion of the Pleistocene river sediments on the west bank. Consequently, the drill hole is not at the top of the sequence. The elevation of the site where the hole was drilled is 2657 feet. Equisetum occurred in the Mojave River in the 19th century, and may exist there today. However, fossils of Equisetum have never been reported from Mojave River sediments or from the lacustrine? Manix Formation. Furthermore, the only record in California of Pleistocene Equisetum we have been able to find is from San Bruno (San Mateo County) and is in the UCMP collection. The San Bruno flora was dated at 10,170 radiocarbon years BP. It is surprising that Equisetum has not been identified in collections from Rancho La Brea. Those collections have produced fish such as stickleback and trout. There certainly was sufficient aquatic habitat to support Equisetum.

Equisetum habitat is freshwater wetlands or shallow waters. Equisetum stems are more likely to be preserved in some freshwater sediments than many aquatic plants because their composition is 10-25% silica. Equisetum fossils seem to be more common in marine rocks than in fresh water sediments. The first author has seen many specimens in shale facies of the Miocene Topanga Formation and Monterey Formation (= Puente Formation), but not in diatomaceous facies. The Equisetum stems and angiosperm leaves often washed out from river mouths into the sea. The marine shale sediments where they are found are typically finer grained than most Neogene freshwater sediments in California, and this may explain the discrepancy. The marine sediments also crop out over a greater area than Neogene freshwater sediments.

No vertebrate fossils were detected during boring or through sediment sampling. Numerous fecal pellets of an unidentified arthropod were found in an exposed fluvial carbonate layer. They may belong to dragonfly larvae or crayfish. Both the fecal pellets and the Equisetum specimens have been accessioned by the San Bernardino County Museum.

This discovery demonstrates the importance of monitoring borings for paleontological resources.

**Eocene fluvial deposits, paleogeography, and basin geometry of the Eocene Goler Formation, Mojave Desert, California**

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The Goler Formation consists of at least 2500 m of continuous fluvial strata deposited just southeast of the southern margin of the Sierra Nevada. We re-evaluated the lower half (1254 m) of the Goler Formation beginning at the basal unconformity in the central part of the basin. Detailed sedimentologic observations are combined with new paleocurrent data (imbricated clasts, epsilon crossbeds, n=64) to document the vertical changes in stratigraphic architecture. Five lithofacies define 3 map units (1-3) based on the relative percentage of each lithofacies and architectural elements. Unit 1 spans from 0–175 m and is dominated by 0.5–2.0 m thick, laterally continuous, immature sandstone beds with minor siltstone interbeds deposited unconformably on the El Paso Mountains basement rocks. These strata are interpreted as sand flat and sheet-flow deposits, with N-directed epsilon crossbeds (018°, 003°; n=3, 5). Unit 2 spans from 175-629 m and is dominated by siltstone with occasional, laterally discontinuous 1–2 m thick sandstone beds. These strata are interpreted as sand and mud flat deposits with minor meandering channel systems prograding across into the mud flat. Unit 3 spans from 629–1254 m and contains 10–20 m thick sections of interbedded sandstone and conglomerate alternating with 10–20 m thick sections of siltstone and shale. Petrified logs and laminated shale are common. Conglomerates provide NNE-directed paleocurrents from imbricated clasts (036°, 354°; n=23, 13). This section is interpreted as a northerly meandering river system with a main trunk channel and overbank components. In summary, early deposition of the lower Goler Formation in our section was locally derived in low-energy fluvial systems confined to an isolated valley floor. The distinct change to a larger...
meandering river system at 629 m may reflect a Paleocene or Eocene drainage capture and basin-amalgamation event that linked the Goler basin to surrounding basins. This large new river must have linked the ancient Goler Basin with the Pacific Ocean to the west, and provided a conduit for the marine incursion that is observed at the top of the Goler Fm. Although the cause of the drainage capture event remains speculative, we suspect it may have been related to regional subsidence that occurred after subduction stopped along this portion of the Cordilleran margin.

Preliminary tephrochronology for the upper Miocene Powerline sequence of wetland deposits, Mojave Desert, CA

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A broad expanse of Miocene to Pliocene sandstone is exposed near the southern margin of the Fort Irwin National Training Center from Alvord Mountain to the Cronese Lakes. These deposits were previously mapped simply as partially consolidated materials, but our mapping has shown that these deposits are predominantly fluvial and can be subdivided although they lack strong age control. An approximately 220-meter-thick sedimentary sequence in this area has characteristics of wetland deposits in the form of abundant calcium carbonate as nodules and porous matrix, abundant rhizoliths, and opal beds interbedded with fine sands and tephra. These wetland deposits are overlain by the Bicycle Lake basalt flow that formed the southern margin of a larger volcanic field. We informally refer to these wetland facies as the Powerline sequence.

Numerous tephra layers found throughout the sedimentary sequence provide an opportunity to improve age control for the sequence and capping basalt flow. The latter has been variably dated at ca. 5.6 Ma, ca. 3.4 Ma and, most recently, 4.55 ± 0.07 Ma.

Thirty-six tephra samples were collected from at least eleven and possibly as many as fourteen distinct tephra layers in a 2.5 square-kilometer section of the Powerline sequence. Samples were processed and analyzed for major and minor elemental concentrations via electron microprobe (EMP). We determined tentative geochronologic constraints by correlating our geochemical results to previously identified tephra in the USGS Tephrochronology Project’s reference database using standard tephrochronology methods.

Three tephra beds provide a general geochronologic framework in the basin. The glass shard composition from a tephra bed near the base of the Powerline sequence correlates well with either the ca. 9.3 Ma McGuire Peak ash bed (Snake River Plain source) or 8.9–8.6 Ma MOD-3/19 ash bed (proto-Cascade source) in the marine Modelo Formation near Ventura, CA. A different tephra layer located nearer to the top of the section shows equally strong correlations with the 6.27±0.04-Ma Walcott tuff, 6.7±0.1-Ma Blacktail Creek ash, and 7.0±0.1-Ma Cub River ash from the Heise volcanic field of the Snake River Plain. These three eruptions are geochemically indistinguishable from one another at the present level of analysis, so together they broadly constrain this layer in the Powerline sequence to ca. 7.1 to 6.27 Ma. The youngest tephra in the section has similarities to several reference tephra allowing for an age estimate of ca 7 to 3.3 Ma. We anticipate that further tephra analysis will refine the age of the Powerline sequence.

Late Miocene, non-marine sedimentary sequences are widely spaced across the Mojave Desert. Tephrochronology of the Powerline sequence will allow paleogeographic reconstruction of the Mojave province during the late Miocene and correlation to marine sequences to the west.

Lava Bed Mountains mining district, San Bernardino County, California

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The Lava Bed Mountains Mining District encompasses the Lava Bed Mountains, and lies southeast of the Newberry Mountains, west of the northern Bullion Mountains, and northeast of the San Bernardino Mountains in the central Mojave Desert. The northern Lava Bed Mountains are underlain by Tertiary sedimentary and volcanic rocks and Mesozoic granitic rocks, as mapped by T.W. Dibblee, Jr. Three Quaternary basalt lava fields lie nearby. Miocene igneous rocks have instructive intrusive relationships exposed in Sections 24 through 35 of T.7N, R.5E, SBM. In this area, Tertiary dacite porphyry has invaded older quartz monzonite along a structural grain that results in remnant dike-like masses of quartz monzonite with the dacite porphyry to the east and dikes of dacite porphyry in the quartz monzonite to the west. Along this same structural grain, Tertiary andesite porphyry invaded the dacite porphyry to the east and quartz monzonite to the west.

The central part of the Lava Mountains, between the Calico and Emerson Faults, is mainly biotite quartz monzonite with mafic dikes described above. The southern part of the Lava Mountains, southwest of the Emerson Fault, are underlain by the Quartz Monzonite of Emerson Lake and biotite diorite.

The Northern Bullion-Lava Bed Mountains area of the mining district can be divided into 8 blocks by major faults which trend NNW–SSW. Geologic map comparisons and compilations indicate that these fault blocks appear to have sets of secondary shear faults within them, most of which have right lateral separation. Tension gashes form between the secondary shear faults and are loci for the formation of vein deposits.

Mineral commodities of the Lava Bed Mountains Mining District are summarized in the table. A geologic
Geology and history of the Cady Mountains mining district, San Bernardino County, California

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The Cady Mountains are located in the central Mojave Desert east of Barstow and south of Afton Canyon. The oldest rocks of the Cady Mountains are metamorphic rocks of Paleozoic and Mesozoic age that crop out in the northern part of the mountains. The metamorphic rocks were intruded by granite in the north Cady Mountains, by quartz monzonite in the south-central Cady Mountains and by biotite quartz diorite in the eastern Cady Mountains. During Oligocene (?) to Pliocene time volcanic and sedimentary rocks assigned to the Barstow, Yermo, and Newberry formations accumulated.

In the Quaternary, sediments of pluvial Lake Manix formed and now crop out northwest of the Cady Mountains. There are 23 mineral commodity groups in the Cady Mountains. They are summarized in the table below.

The Cady mountains were likely prospected during the 1880s to 1920s when large mines were discovered and developed in the nearby Lava Bed, North Bullion, Cave, and Cronese mountains. Celestite and crushed stone (jasper, chalcedony) were developed in 1916. Manganese mining commenced at the Lee Yim mine in 1917, at the Big Reef-Black Butte mine in 1918, and at the Afton Canyon Manganese mine by 1919. In 1918 aluminum and several other elements along with silica were developed at the Big Reef Mine. The Afton Canyon fluorite mines began production in 1921. In 1930 barite (Hansen) and phosphate (Lee Yim) were mined. Hectorite was mined at Hector starting in 1931. In 1935 uranium and rare earth elements were discovered in the Hoerner-Ross pegmatite. Lead was discovered at the Preston mine in the early 1930s. Sand and gravel was produced from the Basin and Beecher pits for Southern Pacific railroad

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report with links to regional and local maps and mine catalogues for the Lava Bed Mountains Mining District is available at: https://www.academia.edu/51061630/Lava_Bed_Mountains_Mining_District_San_Bernardino_County_California or http://www.greggwilkerson.com/lava-bed-mountains.html
prior to 1953. Prior to 1958, pumice (Farway, Heather), copper (Old Dominion), clay (Stacsite), gold (Overlook), and limestone (Baxter) were discovered. The hectorite deposits in the Lava Mountains are still being mined. In the Cady Mountains some borrow pits are active. Other mines are idle.

A detailed report for the Cady Mountains Mining District is available at https://www.academia.edu/44076044/Geology_and_Mining_History_of_the_Cady_Mountains_San_Bernardino_County_California and at http://www.greggwilkerson.com/cady-mountain.html

### Northern Bullion Mountains mining district, San Bernardino County, California

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The northern Bullion Mountains mining district is southwest of the Bristol Mountains, west of the Lava Hills, northwest of the southern Bullion Mountains, east and southeast of the Lava Bed Mountains, and northeast of the Hidalgo Mountains.

The oldest and longest-producing mine of the district is the Bagdad-Chase copper mine, which operated from 1904 to 1976. Bently Resources conducted an exploration

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<td>Hector Mines and Deposits</td>
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<td>Old Dominion, Cu prospect #7</td>
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<td>Secondary copper minerals with minor chalcopyrite in fault zone</td>
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<td>Clay</td>
<td>Stacsite Mine, Clay No. 3</td>
<td>2</td>
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<tr>
<td>Gold</td>
<td>Unnamed Au Producer #2, Au-Cu Producer #1, Overlook</td>
<td>6</td>
<td>pre-1958</td>
<td>Fissure vein and breccia deposits of quartz with associated lead and copper sulfides and copper carbonate minerals hosted by granitic or metavolcanic rocks</td>
</tr>
</tbody>
</table>
program in 1987; 20,000 troy oz of gold were identified but the mine was not reopened. Most of the other large mines were gold producers that were active from 1933 to 1953.

The bedrock units of the northern Bullion Mountains mainly are Tertiary volcanic intrusive and extrusive basalt, andesite, and andesite porphyry, with interbedded pyroclastic (tuff breccia, ryholitic felsite) and sedimentary (fanglomerate, tuffaceous conglomerate) units. There is a small area of Mesozoic quartz monzonite in the southeastern part of the district east of Black Top Mountain. Mineralization and ore body development are structurally controlled.

Comparison and compilation of geologic mapping since 1955 of the northern Bullion Mountains reveals that the area can be divided into 4 blocks by primary strike-slip faults that strike NNW. These fault blocks have sets of secondary faults within them that strike north. From the Rodman Mountains across the northern Bullion Mountains, the blocks are:

Fault Block I is 3.9 miles wide. It is between the Calico Fault to the west and the Bullion Fault to the east. The 4407 Hill Fault is a secondary fault within this block.

Fault Block II is 4.2 miles wide. It is between the Bullion Fault to the west and the 3507 Hill Fault to the east. Between these shears are secondary faults: the Pisgah Fault is 15.96 miles long, Fault A is 8.76 miles long, and Fault AA is 6.10 miles long. The northern part of the Pisgah Fault connects with the 3507 Hill Fault near the Inerto and National Lead mines. The southern part of the Pisgah Fault is a secondary fault within Block I.

Fault Block III is 5.2 miles wide. It extends from the west side of the Lava Bed mountains to the southwest across the Bullion Mountains to the northeast. It is between the 3507 Fault to the west and the Bullion Range Mine Fault to the east. Between these faults is secondary fault B. It can be traced on the surface for 11.39 miles.

Fault Block IV is 5.1 mile wide. It is between the Bullion Range Mine Fault to the west and the Ludlow Fault to the east. It extends to the northeast side of the Bullion Mountains. This block contains most of the productive copper and gold mines in the northern Bullion Range. They follow the Bagdad fault lineament which is a secondary fault that can be traced on the surface for 2.47 miles. The mines along this lineament exhibit common brecciation and ore development under a layer of volcanic rock with shallow dip to the southeast. Within this block is Fault C which is 10.7 miles long. It truncates the Old Pete Fault which can be traced on the surface for 1.25 miles. Faults E (2.03 miles) and D (2.37 miles) connect the Bullion Range Mine Fault with Fault C.

Fault Block V is 8.5 miles wide. It is between the Ludlow Fault to the west and Broadwell Lake Fault (Dokka, 1980) to the east. It encompasses a broad valley and the Lava Hills. Between these faults is the Ash Hill–Klondike–Siberia Fault. There are six mines on or near the Ludlow Fault: Dollar claims (copper); Unnamed (barite); Unnamed prospect (barite); New Process mine (gold); Mineral Monarch (copper); Unnamed barite occurrence (barite).

The commodity distribution for the mines in the northern Bullion Mountains mining district is summarized in the table above.

A common feature of most of the mineral deposits in the north Bullion Mountains is that they are associated with breccias. The USGS (MRDS, 2011) listing for the Bagdad-Chase mines categorizes them as detachment breccias. Polovina (1980a, 1980b) describes the breccias as being related to intrusion of sill solutions which gave rise to hydrothermal breccias. Minedat (2020) describes copper porphyry type alteration in the district. Collier

<table>
<thead>
<tr>
<th>Commodity</th>
<th>Number of deposits</th>
<th>Deposit names</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Barium-Heathite</td>
<td>5</td>
<td>Unnamed prospects</td>
<td>Most are on or near the Ludlow Fault</td>
</tr>
<tr>
<td>Bentonite</td>
<td>1</td>
<td>Unnamed bentonite prospect group</td>
<td>Hectorite</td>
</tr>
<tr>
<td>Bismuth</td>
<td>1</td>
<td>Bagdad - Chase Mine</td>
<td></td>
</tr>
<tr>
<td>Clay</td>
<td>3</td>
<td>Unnamed clay occurrences</td>
<td></td>
</tr>
<tr>
<td>Copper</td>
<td>6</td>
<td>Dollar claims, Mineral Monarch, Bagdad-Chase Mine, Unknown prospects</td>
<td>Most on Bagdad-Chase lineament associated with breccias</td>
</tr>
<tr>
<td>Gemstone</td>
<td>1</td>
<td>Dictator (opal)</td>
<td>Opal in vesicular basalt</td>
</tr>
<tr>
<td>Molybdenum</td>
<td>2</td>
<td>Imperial Lode</td>
<td>Molybdenum (wulfenite)</td>
</tr>
<tr>
<td>Sand and Gravel, Construction</td>
<td>1</td>
<td>Unnamed location</td>
<td>Most on Bagdad-Chase lineament associated with breccias</td>
</tr>
<tr>
<td>Silver</td>
<td>5</td>
<td>Ambush Canyon Mine, Stedman mill site ruins, Unnamed prospects</td>
<td>Most on Bagdad-Chase lineament associated with breccias</td>
</tr>
</tbody>
</table>
(1957) describes some of the deposits as hydrothermal replacements in pyroclastic rocks. None of these interpretations explain the concentration of mines along the Bagdad-Chase lineament. My interpretation is that this geometric orebody association is best understood as the result of a fortuitous geologic history of low angle faulting and brecciation that was subsequently partitioned by high-angle faults associated with right lateral regional deformation. These shear stresses produced local areas of low pressure within the secondary shear structures into which ore fluids migrated, cooled, differentiated, and crystallized to form the ore deposits subsequent to ground preparation.

A report for the northern Bullion Mountains with regional, local, and mine maps as well as mine catalogs can be accessed at https://www.academia.edu/51070843/Northern_Bullion_Mountains_Mining_District_San_Bernardino_County_California or http://www.greggwilkerson.com/northern-bullion-mountains.html

What? There’s water in the desert?*
Carole Ziegler
gologycarole@gmail.com

In 2014, California passed the Sustainable Groundwater Management Act and listed Borrego Spring’s sole-source aquifer to be in “critical overdraft”. With that in mind, this talk was developed for the general audience that visits the Imperial Valley Desert Museum, located near Highway 8 in Ocotillo, California. The talk looks at the sources of water in the Borrego Valley Ground Water Basin including rain storms, lakes, springs and aquifers. Of particular interest is the Borrego Springs Aquifer which occurs as 2 sub-basins, one near the town of Borrego Springs and the other in the area of Ocotillo Wells. Besides being divided into sub-basins, the aquifer’s waters exhibit 3 distinct layers associated with age. Due to natural geologic barriers, waters from one sub-basin cannot be pumped to the other sub-basin. Secondly, there is not enough rainfall available for recharge of these sub-basins. Borrego Springs is currently developing a plan to reduce water consumption by almost 75% by 2040 as required by the state of California. If they don’t, the state will take over and manage the water reduction themselves.

* first published in the 2020 Desert Symposium volume

Detail of the Broadwell Mesa basalt, photo by D.M. Miller, 2013
Robert E. Reynolds
Desert Symposium
Student Research Award

This award to honor Bob Reynolds acknowledges Bob’s decades of service to desert sciences, from directing large fossil excavations and exploring for minerals to mentoring numerous students and apprentices. For these and other achievements, Bob received the 2019 Morris Skinner Prize from the Society of Vertebrate Paleontology. In addition, Bob has been central to holding the annual Desert Symposium for over 30 years, in many cases singlehandedly soliciting contributors, organizing the meeting, and running the field trip. Bob’s leadership and service are honored with this award by promoting student research projects.

Information on applying for and donating to the award is available at http://desertsymposium.org. Donors will be identified in the annual volume published by the Desert Symposium. Desert Symposium Inc. is a non-profit 501(c)3 organization. Contributions are tax-deductible as allowed by law.

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2022 recipient: Kimberly Schaefer, Claremont Graduate University & California Botanic Garden, for the research proposal: “A Vascular Flora of the Sacatar Trail Wilderness”

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