Meteoric fluid infiltration in crustal-scale normal fault systems as indicated by δ¹⁸O and δ²H geochemistry and ⁴⁰Ar/³⁹Ar dating of neoformed clays in brittle fault rocks

Samuel Haines¹*, Erin Lynch¹, Andreas Mulch²,³, John W. Valley⁴, and Ben van der Pluijm¹

¹UNIVERSITY OF MICHIGAN, DEPARTMENT OF EARTH AND ENVIRONMENTAL SCIENCES, 1100 N. UNIVERSITY AVENUE, ANN ARBOR, MICHIGAN 48109, USA
²SENCKENBERG BIODIVERSITY AND CLIMATE RESEARCH CENTER, SENCKENBERGLAGE 25, 60325 FRANKFURT, GERMANY
³INSTITUTE OF GEOSCIENCES, GOETHE UNIVERSITY FRANKFURT, ALTENHOFERALLEE 1, 60438 FRANKFURT, GERMANY
⁴DEPARTMENT OF GEOSCIENCE, UNIVERSITY OF WISCONSIN–MADISON, MADISON, WISCONSIN 53706, USA

ABSTRACT

Both the sources and pathways of fluid circulation are key factors to understanding the evolution of low-angle normal fault (LANF) systems and the distribution of mineral deposits in the upper crust. In recent years, several reports have shown the presence of meteoric waters in mylonitic LANF systems at mid-crustal conditions. However, a mechanism for meteoric water infiltration to these mid-crustal depths is not well understood. Here we report paired δ¹⁸O and δ²H isotopic values from dated, neoformed clays in fault gouge in major detachments of the southwest United States. These isotopic values demonstrate that brittle fault rocks formed from exchange with pristine to weakly evolved meteoric waters at multiple depths along the detachment. ⁴⁰Ar/³⁹Ar dating of these same neoformed clays constrains the Pliocene ages of fault-gouge formation in the Death Valley area. The infiltration of ancient meteoric fluids to multiple depths in LANFs indicates that crustal-scale normal fault systems are highly permeable on geologic timescales and that they are conduits for efficient, coupled flow of surface fluids to depths of the brittle-plastic transition.

1.0 INTRODUCTION

Fluid flow in both individual faults and sets of faults in a given tectonic regime has been the subject of considerable interest for the past 30 years (e.g., Kerrich et al., 1984; McCaig, 1997; Gébelin et al., 2012; Menzies et al., 2014). Fluids in middle- and upper-crustal normal faults show a strong influence of variably evolved, meteoric-derived fluids (e.g., Fricke et al., 1992; Mulch et al., 2004; Swanson et al., 2012; Hetzel et al., 2015). These observations would require downward circulation of surface waters into the mid-crust, a physiomechanical process that is poorly understood (Connolly and Podladchikov, 2004; Person et al., 2007; Lyubetskaya and Ague, 2009). In addition to fluid pathways, fault zone minerals with a meteoric fluid origin can be used to make inferences about regional paleoelevations (e.g., Mulch et al., 2004; Gébelin et al., 2012, 2013). By contrast with normal fault systems, it is thought that fluids in thrust faults are dominated by upward circulation of deep basinal fluids, with minor contributions from evolved meteoric fluids in late stages of orogeny (e.g., McCaig et al., 1995; Trave et al., 2007; Sample, 2010).

Fluid-flow models have outlined a set of narrow permeability and topography conditions by which downward flow of meteoric-water–dominated waters might still occur (Person et al., 2007). Key to testing the feasibility of these fluid-flow models for LANFs is isotopic data from the upper and middle reaches of LANF systems. Isotopic studies of the upper and middle reaches of LANF systems, which extend from the surface to the middle crust, are few relative to the now data-rich mylonitic rocks. Stable isotopes have been employed in faults in carbonate-dominated sequences (e.g., Losh, 1997; Losh et al., 2005; Swanson et al., 2012), but relatively few LANFs occur in carbonate-dominated sequences relative to those in silicate-dominated upper-crustal sections.

Neofomed clay-rich fault gouges are a common feature of LANFs (e.g., Haines and van der Pluijm, 2012) and have been recognized to both dramatically reduce the frictional strength of fault zones (e.g., Carpenter et al., 2011; Haines et al., 2014) and document the age at which fault gouge formation occurred (Solum et al., 2005; Haines and van der Pluijm, 2008). The clay minerals that are neofomed in LANF gouge thus have a major influence on fault behavior, but their potential as recorders of upper-crustal fluid circulation in LANFs has not been broadly examined to date. Phyllosilicates are unusual silicate minerals in that they contain structural hydrogen in addition to the oxygen that is found in all silicates and thus permit analysis of both δ¹⁸O and δ²H on a single mineral phase, allowing for a more complete characterization of the exchanging fluid. The δ²H value of the clay minerals preserves the initial source of the fluid until water-rock ratios become very low (waterrock <0.001; Menzies et al., 2014). By contrast, because all silicate minerals contain oxygen, the δ¹⁸O value is strongly sensitive to the degree of wall-rock–fluid interaction (Sharp, 2005). An analysis of both isotopic ratios from the same mineral separate allows for an evaluation of both the initial source and any degree of wall-rock–fluid interaction of fluids exchanging with that mineral phase.

*Present address: School of Earth Sciences, The Ohio State University, Columbus, Ohio 43210, USA; haines.278@osu.edu

**http://orcid.org/0000-0003-0167-5336
 Fluids with δ²H < -80‰ and δ¹⁸O < 0‰ are generally interpreted to be of meteoric origin, while fluids with δ¹⁸O > +5‰ and δ²H > -80‰ are interpreted to be of metamorphic or igneous origin (Sheppard, 1986).

Illitic clays are common to many clay-rich gouges; they contain K, and retain Ar, permitting dating of clay growth in gouge by ⁴⁰Ar/³⁹Ar methods. We use the illite age analysis method, which utilizes ⁴⁰Ar/³⁹Ar dating in conjunction with quantitative X-ray diffraction (XRD) to determine the age of authigenic and detrital (cataclastically derived) clay mineral populations.

To investigate the fluid-flow system of LANFs in the U.S. Basin and Range province as a class of fault, we utilized a suite of samples that were characterized as part of a companion study of clay gouge mineralogy (Haines and van der Pluijm, 2012), which identified systematic patterns of clay mineral transformations in clay-rich fault gouges. These faults range from shallow-rooted structures (such as the Panamint Range–Front detachment) to LANFs that reached mid-crustal depths (e.g., the Ruby Mountains detachment). We isolated authigenic phyllosilicate minerals from both upper-crustal clay-rich fault gouges and mid-crustal metasomatic, chlorite-rich breccias from a suite of faults (Fig. 1) and analyzed δ¹⁸O and δ²H values of neoformed phases in order to investigate the composition of fluids from which they grew. A subset of these samples was analyzed to determine the age of neoformed clays, and thus, the timing of fluid infiltration.

2.0 LOW-ANGLE NORMAL FAULTS AND FAULTS SAMPLED

Low-angle normal faults (LANFs) are a special class of normal fault, first noted in the American Cordillera (Anderson, 1971; Wernicke, 1981) and now recognized globally (e.g., Collettini, 2011). These faults are unusual in that they have accommodated normal displacements of tens of kilometers and many slipped at dips below those predicted from conventional rock friction arguments (Axen, 2004; Haines and van der Pluijm, 2010). Many exposures of exhumed shallow-crustal LANFs have well-developed cm-thick to m-thick, clay-rich fault gouges that are dominated by neoformed clay minerals, predominantly illite, illite-smectite, and smectite. These neo mineralized clays in fault gouge comprise the uppermost part of a suite of distinctive fault-related rocks in metamorphic core complexes (MCCs) that record progressive exhumation of footwall lithologies, often from pre-faulting mid-crustal depths. Many (but not all) LANF footwall exposures have clay gouges in direct contact with
a distinctive greenschist-facies epidote + chlorite alteration of footwall metamorphic or igneous lithologies (Fig. 2A). This distinctive epidote + chlorite alteration can extend for tens of meters into the footwall, and, where brecciated, these rocks are lithified cataclasites, sometimes called “chlorite microbreccias” (Phillips, 1982; Selverstone et al., 2012). The fault rocks that are inferred to form at greatest depths are commonly quartzofeldspathic mylonites (formed at temperatures from 400 to 550 °C; Anderson, 1988; Mulch et al., 2007).

Fluid flow in LANF systems has been examined with numerous isotopic studies of the mid-crustal mylonitic fault rocks using both δ18O and δ2H on minerals and fluid inclusions (Lee et al., 1984; Wickham and Peters, 1990; Fricke et al., 1992; Wickham et al., 1993; Peters and Wickham, 1995; Mulch et al., 2004, 2007; Gébelin et al., 2011, 2012, 2015; Gottardi et al., 2011). The greenschist-facies microbreccias have also been studied, often in conjunction with the higher-temperature mylonites (Kerrich and Hyndman, 1986; Kerrich and Rehrig, 1987; Kerrich, 1988; Smith et al., 1991; Morrison, 1994; Nesbitt and Muehlenbachs, 1995; Morrison and Anderson, 1998). While many of these studies used δ18O and δ2H analyses, few performed both analyses on the same phase. Fluid circulation in LANFs in carbonate-dominated successions has been studied using carbonate veins formed in lower-temperature (30–300 °C) fault rocks (Losh, 1997; Losh et al., 2005; Swanson et al., 2012). The majority of these studies have documented low-δH/low-δ18O fluids, inferred to be of meteoric origin, although some (Smith et al., 1991) have documented predominantly igneous-dominated fluids, or the interaction of two (metamorphic and meteoric) fluid sources (Kerrich, 1988).

In recent years, δH isotopic studies of neoformed micas in LANF mylonites have yielded very depleted (δH < -100‰) values, interpreted to be indicative of (1) high-altitude meteoric fluid, (2) high-latitude meteoric fluid, or (3) a paleo-rain shadow (Mulch et al., 2004, 2007; Gottardi et al., 2011; Gébelin et al., 2011, 2012, 2015). The processes and pathways by which meteoric fluids of surface origin reach the mid-crust are controversial. Fluids migrating down a fault system will encounter unfavorable thermal and density gradients, and the buoyancy of hot waters at higher pressures is greater than that of colder waters, inhibiting downward flow (Connolly and Podladchikov, 2004; Lyubetskaya and Ague, 2009). In addition, mid-crustal rocks are widely assumed to lack the porosity and permeability to permit fluid flow at rates sufficient to prevent the very low water-rock ratios that would obscure the initial source of the fluid. Some studies, therefore, have suggested that isotopic evidence for meteoric fluids in mid-crustal lithologies is instead evidence of burial of pre-metamorphic fault rocks to mid-crustal depths and not actual incursion of meteoric fluids to mid-crustal shear zones (Clark et al., 2006; Raimondo et al., 2011, 2013). Person et al. (2007) presented a numerical model that suggested that a metamorphic core complex with a fracture-dominated flow system with a relatively narrow range of effective fault
and 380–420 °C, Selverstone et al., 2012) by an influx of infiltrating fluids, providing information on fluid sources and pathways.

2.3 Detachments Sampled

We sampled gouges from five suites of Cordilleran LANFs, comprising eight separate detachments: (1) The Ruby Mountains detachment in northern Nevada; (2) three detachments in Death Valley, California (Badwater detachment, Mormon Point detachment, and Amargosa detachment); (3) two detachments in the Panamint Mountains, west of Death Valley (Panamint Range–Front LANF and Mosaic Canyon detachment); (4) the Buckskin-Rawhide detachment in NW Arizona; and (5) the Waterman Hills detachment in southern California (Figs. 1 and 2). Further description of sampled outcrops is given in Data Repository File DR1; geospatial data in .kmz format are found in Data Repository File DR2. Gouges from these faults were all mineralogically characterized as part of a previous study that identified systematic patterns of clay mineral neomineralization in clay-rich fault gouges (Haines and van der Pluijm, 2012), and the samples analyzed in this study are all a subsample of samples from that study.

3.0 SAMPLE PREPARATION AND CHARACTERIZATION

Fault gouges are mixtures of fragmental wall-rock material derived from one or both sides of a fault zone and authigenic (neformed) clay minerals growing in the gouge. Isolating the neoformed clay component of clay-rich gouges is therefore a challenging process, because the clay crystallites are very small (<<2.0 μm). Gouges can also contain fragmental phyllosilicates that are superficially similar to the authigenic phases but would contaminate the isotopic value without careful characterization. Our sampling approach is shown in Figure 3. We use gravity settling in water to isolate the <2.0 μm (Stokes equivalent diameter) size fraction, followed by high-speed centrifugation to separate the clay-size fraction into three or four size fractions, coarse (2.0–7.0 μm), medium (0.02–0.05 μm), and fine (<0.05 μm). Each fraction is then characterized by XRD, using both oriented mounts (with and without ethylene glycol solvation) to identify principal clay phases and random powder mounts to accentuate the non-(001) peaks characteristic of clay polytypes (which can be used to identify authigenic clay minerals in gouges). Additional site information and mineralogical description of these samples are found in Data Repository Files DR1 and DR2 and Haines and van der Pluijm (2012).

3.1 Sampling Clay-Rich Gouges

We analyzed only gouge clay samples that were well characterized in previous studies (Haines and van der Pluijm, 2010, 2012) for this study with >90% authigenic material based on XRD. X-ray diffraction patterns of all analyzed materials are given in Figure 4. Isotopic measurements of clay minerals in these samples are reported in Table 1. For our sampling approach, we use gravity settling in water to isolate the <2.0 μm (Stokes equivalent diameter) size fraction, followed by high-speed centrifugation to separate the clay-size fraction into three or four size fractions, coarse (2.0–7.0 μm), medium (0.02–0.05 μm), and fine (<0.05 μm). Each fraction is then characterized by XRD, using both oriented mounts (with and without ethylene glycol solvation) to identify principal clay phases and random powder mounts to accentuate the non-(001) peaks characteristic of clay polytypes (which can be used to identify authigenic clay minerals in gouges). Additional site information and mineralogical description of these samples are found in Data Repository Files DR1 and DR2 and Haines and van der Pluijm (2012).

3.2 Geochemical Methods

Isotopic measurements of clay minerals in these samples are reported in Table 1. For our sampling approach, we use gravity settling in water to isolate the <2.0 μm (Stokes equivalent diameter) size fraction, followed by high-speed centrifugation to separate the clay-size fraction into three or four size fractions, coarse (2.0–7.0 μm), medium (0.02–0.05 μm), and fine (<0.05 μm). Each fraction is then characterized by XRD, using both oriented mounts (with and without ethylene glycol solvation) to identify principal clay phases and random powder mounts to accentuate the non-(001) peaks characteristic of clay polytypes (which can be used to identify authigenic clay minerals in gouges). Additional site information and mineralogical description of these samples are found in Data Repository Files DR1 and DR2 and Haines and van der Pluijm (2012).
Figure 3. Illustration of sample preparation process. (A) Field photograph of gouge sampling locality (Badwater-1). Sample Bad-1 (F) is illitic gouge from the pictured gouge layer. Sample Bad-1 (G) is the <0.05μm fraction from disaggregated footwall. (B) Schematic representation of fault gouge in situ, highlighting neoformed clays in fault gouge and fragmental minerals, originating from the wall rock. (C) Separation of the clay fraction (<2μm) by settling in water. (D) Centrifugation of the clay fraction into coarse (2.0–0.2μm), medium (0.2–0.05μm, abbreviated “M”), and fine (<0.05μm, abbreviated “F”) size fractions. All size fractions are then characterized by X-ray diffraction (XRD). Only medium- and fine-size fractions of clays that were nearly monomineralic to the level of XRD detection limits (<5%–10% other phase) were analyzed for O and H isotopes in this study.

Figure 4. X-ray diffraction analyses of gouges sampled in this study. (A) Illite-rich separate from gouges where illite is the neoformed mineral. Patterns are collected from random-mounted samples to highlight polytype-specific peaks. Characteristic (hkl) peaks of illite are shown at top. Gray boxes highlight broad humps centered at 24.9° and 29.1° and are characteristic of the low-temperature 1Md polytype of illite. Note that the intensities of 1Md peaks are variable, depending on whether the illite is cis-vacant or trans-vacant, and that samples lacking clear 24.9° and 29.1° peaks are still the 1Md polytype. Q—quartz (present in the ASH-1 sample at near-detection limits). (B) Smectites separated from gouges where smectite is the authigenic clay phase. Patterns are collected from oriented ethylene glycol air-saturated samples to swell smectite interlayers. Characteristic (hkl) peaks of smectite are shown at top. K—kaolinite, present at near-detection limits in WH68-3 (MF). (C) Gouges where chlorite is the dominant clay mineral in the gouge. The chlorite is fragmental, derived from chlorite-epidote microbreccia footwall lithologies, and is not neoformed in the gouge. Patterns are collected from unoriented samples to highlight higher-order (hkI) reflections and polytype-specific peaks. Characteristic (hkI) peaks for chlorite are shown at top. I = 10 Å mica (illite, muscovite, or biotite) present in near-detection-limit quantities in samples A-BOMB 3 (M) and MOR-3 (M). C—calcite, present in near-detection-limit amounts in sample MOR-3 (M). Note that for all samples, (F) in sample name indicates size fraction <0.05 μm (Stokes equivalent); (M) represents 0.2–0.05 μm size fraction; (MF) indicates <0.2 μm size fraction. Note that by contrast, (G) in a sample name refers to the green color from chlorite and does not connote a size fraction. BAD-1 (G) is a <0.2 μm size fraction.
were made on splits from the same material described in Haines and van der Pluijm (2010, 2012). We note that three of the 14 samples contain near-detection-limited quantities of one or two other mineral phases: quartz in ASH-1, a 10-A phase (illite, muscovite, or biotite/phlogopite) in MOR-3 and A-BOMB-3, and calcite in MOR-3. Although all illitic material contained some interlayered smectite as discernable by XRD, for this study we only used illitic clays that were >80% illite/smectite, and most were >90% illite/smectite.

3.2 Sampling Gouges Derived from Epidote/Chlorite Microbreccias

For this study, we only sampled chlorite-rich gouges where no other Mg-rich phyllosilicates (tri-octahedral clays such as saponite, chlorite/smectite, corrensite, talc/stevensenite, or vermiculite-like phases, or other phyllosilicates such as sepiolite or palygorskite) were detectable by XRD. These samples also are free of other phases (e.g., epidote, feldspar, and calcite) to the level of XRD detection (Fig. 4C). In these gouges, chlorite in the fault gouge is structurally and compositionally indistinguishable from that found in chlorite-epidote alteration zones in the fault footwall as determined by XRD (Haines and van der Pluijm, 2012). These purely cataclastic gouges have effectively disaggregated the footwall lithologies, allowing footwall-derived chlorite grains to be efficiently separated by settling in water and subsequent centrifugation, similar to the authigenic clays in clay-rich gouge (see above).

4.0 ANALYTICAL METHODS

4.1 δ18O Isotopic and δ2H Measurements

Oxygen isotopic analysis of clay separates was completed in the University of Wisconsin Stable Isotope Laboratory by laser fluorination using BrF5 (Valley et al., 1995) and an airlock sample chamber that prevented pre-fluorination (Spicuzza et al., 1998). Hydrogen isotopic measurements were made by continuous-flow mass spectrometry at the Stable Isotope Laboratory at Leibniz Universität Hannover, except for samples WH68-1 (F) and WH68-3 (MF) that were analyzed at the U.S. Geological Survey in Denver. All isotopic ratios are reported relative to Vienna standard mean ocean water (VSMOW), and methods are detailed in Data Repository File DR3.

4.2 Illite Age Analysis (IAA)

40Ar/39Ar ages of samples were obtained by vacuum encapsulation (Dong et al., 1995) to address Ar loss during sample irradiation (“Ar recoil”). Samples were packaged into fused silica vials and sealed prior to irradiation (van der Pluijm et al., 2001). Thus, the 39Ar expelled from the crystallites during irradiation is retained for analysis (see van der Pluijm and Hall, 2015, for a full description of the method). The sample vials were broken open, the initial gas was analyzed, and the vials were then step-heated under a defocused laser until sample fusion occurred. Note that the total gas age obtained from the vacuum-encapsulated sample is functionally equivalent to a conventional K-Ar age (Dong et al., 1995).

5.0 RESULTS

Stable isotope values for LANF neoformed illite, smectite, and chlorite are shown in Table 1 and Figures 5A, 5C, and 5E. Individual illite δ18O isotope values range from −2.0‰ SMOW (Ruby Mountains, SEC-4 2) to +11.5‰ (Badwater), and illite δ2H values range from −142‰ for both. The Mormon Point detachment samples—MOR-2 (M), MOR-2 (F), and MOR-3 (M)—all have isotopic values for δ18O of −1.8‰ and −2.0‰, respectively, and δ2H of −142‰ for both. Values of δ18O of +0.58‰ to +8.1‰, and δ2H values fall in a relatively narrow range from −97‰ to −113‰. The Mormon Point detachment samples—MOR-2 (M), MOR-2 (F), and MOR-3 (M)—all have relatively low δ18O carbon values ranging from −8.0‰ to +3.1‰ and δ2H values ranging from −99‰ to −108‰. The Chemehuevi detachment (LOBECK-3 [M]) and Badwater detachment (BAD-1 [G]) samples have δ18O values of +2.5‰ and +4.84‰, respectively, and δ2H values of −106‰ and −113‰, respectively. The Buckskin-Rawhide detachment chlorite sample (A-BOMB-3) shows the highest δ values, with δ18O of +8.1‰ and δ2H of −97‰.

Ages of neoformed clay in selected gouge samples are listed in Table 2, and Ar degassing spectra for each grain-size fraction are included in Data Repository File DR4. We illustrate our results with a sample from the Badwater detachment (Fig. 6). Four size fractions show decreasing...
Figure 5. δ¹⁸O and δD values of neoformed low-angle normal fault (LANF) gouge illite, smectite, and chlorite, together with calculated fluid compositions exchanging with each phase. (A) Fault-gouge illite isotopic values plotted together with isotopic data from other geological environments for reference. Illite geological environment data are compiled from references in Data Repository Files DR5 and DR6. Uncertainties for all δ¹⁸O and δD measurements are within sample marker point size. Fault-gouge illite results: S—Savci fault, Turkey (Isik et al., 2014); NA—North Anatolian fault zone, Turkey (Tonguc Uysal et al., 2006); M—Moab fault, Utah, USA (Solum, 2005). (B) Range of calculated fluid compositions of exchanging fluid for illites. Fluid compositions are calculated using fractionation equations of Sheppard and Gilg (1996) and Capuano (1992). Temperatures are bounded on the lower limit by 50 °C or the meteoric water line and on the upper end at 120 °C. (E) Fault-gouge smectite isotopic values plotted together with isotopic data from smectites from other geological environments. Smectite geological environment data are compiled from references in Data Repository Files DR5 and DR6. (MF) indicates <0.2 μm size fraction. (D) Calculated fluid compositions of exchanging fluid for smectites. Fluid compositions are calculated using fractionation equations of Sheppard and Gilg (1996) and Capuano (1992). Temperatures are bounded on the lower limit by 50 °C or the meteoric water line and on the upper end at 120 °C. (F) Fault-gouge chlorite isotopic values plotted together with isotopic data from other geological environments for reference. Chlorite geological environment data are compiled from references in Data Repository Files DR5 and DR6. Fault or shear zone chlorite results: WO—Walter-Outalpa shear zone, Australia, Alice Springs orogen, Australia, and Argentera massif, France (Clark et al., 2006; Raimondo et al., 2011; Leclere et al., 2014); A—Alpine fault (Menzies et al., 2014); P—Picacho metamorphic core complex (Kerrich and Rehrig, 1987); MP—Monte Perdido thrust, Spain (Lacroix et al., 2012). (F) Calculated fluid compositions of exchanging fluid for chlorites. Fluid compositions are calculated using fractionation equations of Cole and Ripley (1998) and Graham et al. (1987). Because significant uncertainties exist for the magnitude of chlorite-fluid exchange, likely fluid compositions are shown with boxes covering range of uncertainty. Please note: (F) in sample name indicates size fraction <0.05 μm (Stokes equivalent); (M) represents 0.2–0.05 μm size fraction. BAD-1 (G) is a <0.2 μm size fraction.
TABLE 2. ILLITE AGE ANALYSIS RESULTS FOR SAMPLES FROM THIS STUDY

<table>
<thead>
<tr>
<th>Detachment fault</th>
<th>Gouge sample</th>
<th>Size fraction*</th>
<th>% 2M1</th>
<th>% 0% 2M1</th>
<th>% 100% 2M1</th>
<th>TGA (Ma) ±</th>
<th>Age (Ma), 0% 2M1</th>
<th>Age (Ma), 100% 2M1</th>
<th>R²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amargosa detachment</td>
<td>Ash-1</td>
<td>2.0–0.2 80%</td>
<td>2.5</td>
<td>139.7 0.2</td>
<td></td>
<td></td>
<td>3.2 ± 3.9</td>
<td>170.4 ± 108</td>
<td>0.998</td>
</tr>
<tr>
<td>Amargosa detachment</td>
<td>Ash-1</td>
<td>0.2–0.05 30%</td>
<td>2.5</td>
<td>49.5 0.17</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Badwater Turtleback</td>
<td>Badwater Turtleback</td>
<td>0.02–0.05 30%</td>
<td>2.5</td>
<td>6.6 2.5</td>
<td>0.2</td>
<td>3.3 ± 0.4</td>
<td>12.2 ± 1.9</td>
<td>0.968</td>
<td></td>
</tr>
<tr>
<td>Badwater Turtleback</td>
<td>Badwater Turtleback</td>
<td>&lt;0.05 2%</td>
<td>2.5</td>
<td>6.6 2.5</td>
<td>0.2</td>
<td>3.3 ± 0.4</td>
<td>12.2 ± 1.9</td>
<td>0.968</td>
<td></td>
</tr>
<tr>
<td>Mormon Point Turtleback</td>
<td>Mormon-1</td>
<td>&lt;0.05 0%</td>
<td>2</td>
<td>2.6 0.45</td>
<td>2.8 ± 0.5</td>
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<tr>
<td>Mosaic Canyon fault</td>
<td>Mosaic-1</td>
<td>2.0–0.2 2%</td>
<td>2.5</td>
<td>49.5 0.17</td>
<td></td>
<td></td>
<td>170.4 ± 108</td>
<td>0.998</td>
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<tr>
<td>Mosaic Canyon fault</td>
<td>Mosaic-1</td>
<td>&lt;0.05 2%</td>
<td>2.5</td>
<td>6.6 2.5</td>
<td>0.2</td>
<td>3.3 ± 0.4</td>
<td>12.2 ± 1.9</td>
<td>0.968</td>
<td></td>
</tr>
<tr>
<td>Panamint Front Range fault</td>
<td>S-Park-1</td>
<td>&lt;0.05 0%</td>
<td>2</td>
<td>3.6 0.17</td>
<td>3.6 ± 0.2</td>
<td></td>
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</table>

Note: TGA—total gas 40Ar/39Ar age. Raw Ar release spectra are available in GSA Data Repository item 4 (see text footnote 1). 2M1—the 2M1 polytype of illite, as ascertained by XRD.

*Stokes equivalent diameter.

Figure 6. Illite age analysis plot and supporting data illustrating clay dating approach. Samples from Badwater detachment are used to illustrate the method. (A1 and A2) Measured (black) and modeled (gray) XRD patterns from fine (<0.05 mm) and coarse (2.0–0.5 mm) fractions respectively. The modeled (gray) XRD spectra quantify the ratio of authigenic (1Md) and detrital (2M1) illite in each grain size fraction. (B1 and B2) Ar release spectra from vacuum-encapsulated material analyzed in A1 and A2. All samples are vacuum encapsulated prior to irradiation to avoid complications associated with Ar recoil during irradiation; total gas ages incorporate both the Ar lost due to recoil (but trapped in the evacuated vial) and retained Ar (see van der Pluijm and Hall, 2014). (C) Illite age analysis plot comparing Ar encapsulation age (total gas age) and % detrital 2M1 polytype of illite in sample. Lower and upper intercepts of York regression on these data constrain the authigenic (3.3 ± 0.4 Ma) and detrital (12.2 ± 1.9 Ma) ages of illite in this gouge sample.

6.0 DISCUSSION

Our results from LANF gouge illites show that they are significantly depleted in both 2H and 18O compared to previously published results from illite formed in three fault gouges from strike-slip and normal fault environments (Fig. 5A; Solum, 2005; Tonguç Uysal et al., 2006; Isik et al., 2014). Because paired oxygen and hydrogen isotopic data from either illite or smectite taken from within fault zones are rare, we compiled published δ18O and δD data for neoformed illite, smectite, and chlorite from several different geological settings to place our results in a broader

percentages of detrital illite with smaller grain sizes (Fig. 6A). Corresponding Ar ages for these samples are systematically younger with decreasing detrital illite, which we analyze in an IAA plot (van der Pluijm et al., 2001; Fig. 6B). Using linear York regression (Mahon, 1996) of percentage detrital illite versus e(−λt) – 1 (where λ is decay constant and t is age) produces extrapolated authigenic and detrital intercept ages of 3.3 ± 0.4 Ma and 12.2 ± 1.9 Ma, respectively. Note that this particular regression analysis treats both parameters as independent, resulting in age errors that primarily reflect the 2%–3% error in mineralogic quantification, while individual Ar ages have much smaller errors, on the order of 0.2–0.5 Ma.
context: (1) sedimentary basins, (2) active and fossil hydrothermal systems, (3) sedimentary basins that experienced meteoric water flushing (illite only, e.g., Whitney and Northrup, 1987), (4) bentonites (smectite only), (5) bentonites where significant post-formation alteration caused $^3$H and $^{18}$O exchange to become decoupled (smectite only, e.g., Cadin et al., 1996; Horton and Chamberlain, 2006), (6) metamorphic rocks (chlorite only), and (7) fault zones (illite and chlorite only). Compiled literature data are shown in Figures 5A, 5C, and 5E. Compiled data are found in Data Repository File DR5, and supporting references in Data Repository File DR6.

6.1 Illite/Smectite Gouge Samples

Our gouge illite samples are isotopically depleted relative to illite that forms in sedimentary basins, and they are also depleted in $^3$H with respect to illite that formed in hydrothermal systems. Some gouge illites (Waterman Hills, Amargosa, and Badwater) have oxygen and hydrogen isotopic compositions that are similar to illites from sedimentary basins interpreted to have been formed during basinal flushing with meteoric water (“meteorically reset”) (Fig. 5A; Glasmann et al., 1989) or hydrothermal systems; but other gouge illites (Ruby Mountains and Panamint) have $^{18}$O and $^2$H values far lower than any reported from sedimentary basins. The Ruby Mountains illite samples preserve hydrogen and oxygen isotope values lower than any illite measurements yet reported ($^{18}$O = $-1.8\%$e and $-2.0\%e$, $^2$H = $-142\%e$ for both). Gouge smectite samples have isotopic compositions that are very similar to illite results, with the Waterman Hills sample (Fig. 5C) and the extremely isotopically depleted Ruby Mountains smectite ($^{18}$O = $+17.9\%e$, $^2$H = $-95\%e$) similar to smectites in sedimentary basins or smectites from bentonites (Fig. 5C) and the extremely isotopically depleted Ruby Mountains smectite ($^{18}$O = $+3.6\%e$, $^2$H = $-147\%e$), which is the most depleted smectite isotopic measurement with respect to both oxygen and hydrogen yet reported.

6.2 Chlorite Microbreccia Samples

The chloride isotopic data are similar to the most isotopically depleted chlorites found in hydrothermal systems (Fig. 5E), with the BAD-1 (G) (M) sample being the lowest $^2$H value yet reported ($^2$H = $-147\%e$). Overall, our chloride samples are very isotopically depleted, especially with respect to hydrogen (all $^2$H = $-97\%e$ to $-113\%e$), relative to those found in metamorphic terranes or in sedimentary basins (Fig. 5E and references in Data Repository File DR6). $^{18}$O values for the chloride samples are more variable, ranging from $+0.6\%e$ to $+11.5\%e$, likely reflecting variable amounts of fluid–rock interaction.

6.3 Equivalent Fluid Compositions

Interpreting stable isotopic values of phyllosilicate minerals and using them to estimate the composition of the fluid with which they exchanged requires constraints on the temperature at which neoformed minerals grew and the associated fractionation between mineral and fluid. While clay gouges lack fluid inclusions that permit direct estimation of the temperature of formational fluids, the clay mineral assemblages found in these gouges place constraints on temperature at their time of formation. Previous studies of clay gouge mineralogy with reliable thermal constraints indicate that neoformed illite in fault gouges from a range of fault settings form at temperatures 80 °C to 180 °C, and perhaps as low as 50 °C (Haines and van der Pluijm, 2012). Because the neoformed illite is the low-temperature 1M$_2$ polytype for all samples and XRD analysis indicates that all samples contain some interlayered smectite, the likely temperature of formation is no more than ~120 °C for both illite and smectite in LANF gouge. From measurements of $^{18}$O and $^2$H and estimates of a plausible clay-formation temperature range, the isotopic composition of the fluid that exchanged with the clay can be calculated. Using published $^{18}$O fractionation equations for illite and smectite (Sheppard and Gilg, 1996), we determine $^{18}$O fluid compositions in equilibrium with the clay phases measured (Figs. 5B and 5D). Similarly, published water-mineral $^2$H fractionation equations for illite and smectite (Capuano, 1992) permit calculation of the fluid composition exchanging with the neoformed clays in LANF fault gouge. Based on this analysis, we find that the compositions of the fluids with which clays exchanged range from nearly pristine meteoric water to weakly isotopically enriched meteoric water. Calculated end-member water compositions are compatible with prior estimates of Middle Miocene (Ruby Mountains and Waterman Hills) and Pliocene (Amargosa, Panamint, and Badwater) Basin and Range meteoric waters (Poage and Chamberlain, 2002; Gebelin et al., 2012, 2015). Only the fluid exchanging with the Badwater gouge illite (Bad-1) shows significant deviation of oxygen enrichment from the field of isotopic values of fluids found in sedimentary basins with increasing depth (Fig. 5B), possibly reflecting oxygen exchange with silicate minerals in the fault zone prior to illite growth. Alternatively, the Death Valley area has been periodically evaporative since the Pliocene (Knott et al., 2005). Evaporative fluids are higher in $^{18}$O than the meteoric water line (Holser, 1979), which might also explain the observed O enrichment of the Badwater sample relative to other samples.

6.4 Previous Fault Zone Isotopic Results

The sole previous oxygen and hydrogen analyses of illite from the gouge of a normal fault, the Moab fault in Utah, USA ($^{18}$O = $+7.9\%e$ and $+8.6\%e$, $^2$H = $-114\%e$ and $-116\%e$, respectively; Solum, 2005) did not report an equivalent fluid composition, but our calculations from the reported mineral values are consistent with a weakly heavy isotope–enriched meteoric fluid ($^{18}$O = $-4.0$ to $-6.5\%e$, $^2$H = $-83$ to $-93\%e$). These limited results support our interpretation of a link between kinematic environment and fluid source, with normal fault systems being dominated by fluids of meteoric origin, while reverse fault systems are dominated by fluids of basinal or metamorphic origin (e.g., Kerrich, 1988; McCaig, 1997). By contrast, data from deeply rooted strike-slip systems suggest that the compositions of the fluids with which clays exchange are more variable in these systems. Data from the crustal-scale North Anatolian fault zone indicate fluid infiltration at various times by fluids of metamorphic or magmatic origin (Tonguç Uysal et al., 2006) and meteoric origin (Boles et al., 2015). Data from the subparallel but shallower-rooted Savcili strike-slip fault zone (Isik et al., 2014) suggest a deep basinal origin for circulating fluids.

The temperature of chloride formation in epistoe-chlorite breccias is less constrained than that for illitic gouges. Estimates range from 300 to 350 °C (Kerrich, 1988) to 350–520 °C (Morrison and Anderson, 1998) and to 380–420 °C (Selverstone et al., 2012). To capture this uncertainty, we use a temperature range of 340–440 °C. The variation in oxygen isotope fractionation over the full range of proposed temperatures (300–520 °C) is $<1.3\%e$, far smaller than the observed range for illite or smectite, and thus the uncertainty in temperature has little effect on interpretation of the chloride data. Hydrogen isotope fractionation between chloride and water is poorly constrained at $-30\%e$ to $-40\%e$ but is thought not to change significantly with temperature over the range at which these breccias formed (Graham et al., 1987). Unlike illitic and smectitic clay minerals, chlorite in both brittle fault zones ($<300 \, ^oC$) and mylonitic green-schist- and amphibolite-facies shear zones has been extensively studied with stable isotopic methods. Previous studies of chlorites in fault zones include LANF (Picacho Mountains metamorphic core complex [MCC], Kerrich and Gehrig, 1987), as well as upper greenschist- and/or lower
amphibolite-facies shear zones in the French Pyrenees (Leclere et al., 2014) and central Australia (Clark et al., 2006; Raimondo et al., 2011) and a Tertiary thrust fault in the Pyrenees active at ~200 °C (Lacroix et al., 2012) (Fig. 5E). The results most germane to this study are chlorite samples from the Picacho Mountains MCC, which have δ18O values of +4.7‰ to +5.5‰ and δD values of −85‰ to −95‰. Our results are similar to these and suggest exchange with a fluid moderately enriched in δ18O but depleted in δD. Overall, chlorites from LANF systems have similar δ18O values to chlorites from amphibolite-facies shear zones inferred to have been infiltrated by meteoric fluids but have far lower δD and δ18O values than chlorites taken from brittle thrusts in compressional tectonic settings (Lacroix et al., 2012). Significantly, our samples all have δD that is ~20‰ lighter than those observed in fault zones other than LANFs.

6.5 Meteoric Water Infiltration and Circulation

Isotopic exchange with fluids of meteoric origin has been increasingly documented associated with faults at mid-crustal depths (Morrison, 1994; Mulch et al., 2004; Gottardi et al., 2011; Gebelin et al., 2012; Mancktelow et al., 2015), but the mechanisms by which surface fluids reach these depths is not well understood (Roddy et al., 1988; Barentt et al., 1996; Losh et al., 2005; Hetzel et al., 2013). Our data from the upper brittle reaches of LANF systems show that meteoric water (with evidence of some wall-rock–fluid interaction) is the predominant fluid in deformed upper crust of LANF systems down to several kilometers depths. Our results showing meteoric fluid infiltration in the brittle portion of LANFs, together with observations of meteoric fluids at greater depths (i.e., chlorite breccias and mylonites) and model predictions, suggest that the drawdown of meteoric water along brittle faults is the dominant fluid circulation system in and near fault zones in extended crust. Convective flow up to balance the fluid-flow system must therefore occur either away from the fault zones or elsewhere up some other reach of the same fault system. Recent studies in the Dixie Valley hydrothermal field have suggested that in some cases, fluids travel updip along discrete sections of basin-bounding normal faults and resurface in hydrothermal springs, the location of which are transient over thousand- to ten-thousand–year timescales, as supported by geochronologic studies of hot spring deposits (Blackwell et al., 2007). Additionally, geothermal modeling of this region suggests that thermal activity and fluid flow along faults may vary according to permeability structure of the fault, with some portions of the fault favoring the upward flow of fluids, whereas other along-strike portions of the fault may behave in a hydraulically opposite sense, allowing fluids to flow downdip (McKenna and Blackwell, 2004; Wanner et al., 2014). Variations in geothermal gradient in the basins also suggest that fluids may flow basinward away from faults through permeable sedimentary layers and layers with favorably oriented fracture networks (Blackwell et al., 2007). Our study of neoformed clays offers novel documentation that supports previous assertions that hanging-wall rocks of evolving LANFs experienced extensive infiltration of surface fluid at least along some, if not all, portions of transient fault and fracture systems to depths of as much as 10 km over time periods of millions of years. This upper-crustal plumbing system provides a pathway for meteoric fluids to mid-crustal depths and formation of mineral deposits by mixing of meteoric fluids with deeper-sourced, metal-enriched fluids (Spencer and Welty, 1986; Roddy et al., 1988).

6.6 Evaluation of Post-Faulting Isotopic Exchange

A concern with stable isotopic analysis of clay minerals is the possibility that the measured isotopic values record late isotopic exchange at near-surface conditions and that the measured values do not reflect the conditions at the time of clay formation at temperatures of 60–180 °C for illite and smectite, or the greenschist-facies conditions at which the cataclastically reworked chlorites originally formed. To address this concern, we: (1) compared our calculated paleofluid compositions to present-day meteoric-water compositions near the faults we sampled, and (2) dated the sample material we used for the stable isotopic measurements by 40Ar/39Ar methods.

The stable isotopic composition of modern precipitation across the western United States has been investigated extensively and was recently reviewed by Lechler and Niemi (2011). The δ18O values of precipitation at sites closest to our sample sites range from −15.6‰ to −8.3‰ with a general trend toward more negative values toward the north and northwest (Friedman et al., 1992, 2002; Lechler and Niemi, 2012; Table 3 and Fig. 7). The δD of precipitation at sites closest to our sample sites range from −115‰ to −57‰, decreasing toward the north and northwest, generally correlating with δ18O and following the global meteoric water line across the Great Basin. The δD values of precipitation do deviate slightly from the meteoric water line during the summer months when evaporative fractionation effects are strongest (Friedman et al., 2002). Our calculated paleofluid compositions record a similar trend in that more isotopically depleted paleofluid compositions are also found at faults where present-day precipitation is strongly isotopically depleted. However, two lines of evidence suggest our calculated paleofluid compositions reflect ancient fluids and not late alteration or mixing with present-day fluids. (1) Calculated fluid compositions are sometimes isotopically heavier with respect to both oxygen and hydrogen (e.g., BAD-1) or lighter (ASH-1, S-PARK-1, WH68-1, and WH68-3) than present-day precipitation (Fig. 7), suggesting that there is not a direct relationship between calculated paleofluid composition and observed present-day precipitation. Where calculated paleofluid compositions are similar to present-day precipitation compositions (SEC 1-2 and SEC 4-2), the required 100–120 °C temperatures are inconsistent with vadose-zone interaction with current precipitation and instead consistent with higher-temperature interaction with an even more isotopically depleted fluid. (2) Dating of the authigenic clays also excludes the concern that mineral isotopic signals are indicative of late, near-surface low-temperature exchange after faulting. All of the 40Ar/39Ar ages for the samples listed in Table 2 are geologically consistent with clay growth while the sampled faults were active. The Ruby Mountains ages (reported in Haines and van der Pluijm, 2010, on splits from the samples used in this study) document the last major period of slip and fluid activity on the detachment at ca. 12 Ma, consistent with previous thermochronometer work (Colgan et al., 2010). The Panamint detachment gouge age of 3.6 ± 0.2 Ma is consistent with an inferred Pliocene time of slip (Andrew and Walker, 2009), while the mid-Miocene age for the Tucki Mountain gouge (16.9 ± 2.4 Ma) is also geologically plausible (Hodges et al., 1990). Likewise, Late Pliocene ages for gouge formation in Armargosa (3.2 ± 3.9 Ma), Mormon (2.8 ± 0.5 ma), and Badwater (3.3 ± 0.4 Ma) detachments of the Black Mountains record the last major pulse of motion on these LANFs (e.g., Knott et al., 2005; Norton, 2011). The dated fault rocks do not show evidence of significant postfaulting alteration, which demonstrates that meteoric fluid signatures preserved in neoformed clays are representative of ancient fluid circulation and not modern surface alteration.

7.0 CONCLUSIONS

Our study of neoformed clays and chlorites in exhumed shallow-crust to mid-crustal LANF systems shows that both LANF clay gouges and mid-crustal chlorite “microbreccias” exchanged isotopically with pristine to weakly evolved meteoric water. The presence of meteoric waters in LANF detachments at multiple crustal levels indicates these systems were hydrologically open for large parts of their history. Instead of recording
TABLE 3. AVERAGE ISOTOPIC VALUES OF PRESENT-DAY PRECIPITATION NEAREST OUR SAMPLING SITES

<table>
<thead>
<tr>
<th>Detachment fault</th>
<th>Range</th>
<th>Gouge sample</th>
<th>Nearest present-day precipitation measurement</th>
<th>Average $\delta^{18}O$ (%) SMOW</th>
<th>Average $\delta^2H$ (‰ SMOW)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amargosa detachment</td>
<td>Black Mountains, California</td>
<td>Ash-1</td>
<td>Death Valley, California</td>
<td>−9.5</td>
<td>−75</td>
<td>Friedman et al. (1992, 2002)</td>
</tr>
<tr>
<td>Badwater turliteback</td>
<td>Black Mountains, California</td>
<td>Bad-1</td>
<td>Death Valley, California</td>
<td>−9.5</td>
<td>−75</td>
<td>Friedman et al. (1992, 2002)</td>
</tr>
<tr>
<td>Mormon Point turliteback</td>
<td>Black Mountains, California</td>
<td>Mormon-2</td>
<td>Death Valley, California</td>
<td>−9.5</td>
<td>−75</td>
<td>Friedman et al. (1992, 2002)</td>
</tr>
<tr>
<td>Mormon Point turliteback</td>
<td>Black Mountains, California</td>
<td>Mormon-3</td>
<td>Death Valley, California</td>
<td>−9.5</td>
<td>−75</td>
<td>Friedman et al. (1992, 2002)</td>
</tr>
<tr>
<td>Buckskin-Rawhide detachment</td>
<td>Buckskin Mountains, Arizona</td>
<td>A-Bomb-3</td>
<td>Parker Dam, Arizona</td>
<td>−8.3</td>
<td>−61</td>
<td>Friedman et al. (1992)</td>
</tr>
<tr>
<td>Waterman Hills detachment</td>
<td>Waterman Hills, California</td>
<td>WH-68</td>
<td>Daggett, California</td>
<td>−10.9</td>
<td>−72</td>
<td>Friedman et al. (1992)</td>
</tr>
<tr>
<td>Waterman Hills detachment</td>
<td>Waterman Hills, California</td>
<td>WH-68-1</td>
<td>Daggett, California</td>
<td>−10.9</td>
<td>−72</td>
<td>Friedman et al. (1992)</td>
</tr>
<tr>
<td>Chemehuevi detachment</td>
<td>Chemehuevi Mountains,</td>
<td>Lobeck-3 (M)</td>
<td>Needles, California</td>
<td>−8.3</td>
<td>−57</td>
<td>Friedman et al. (1992)</td>
</tr>
<tr>
<td>Panamint range front LANF</td>
<td>Panamint Mountains,</td>
<td>S-Park-1</td>
<td>Panamint Range</td>
<td>−13.4</td>
<td>−108</td>
<td>Lechler and Neimi (2012)</td>
</tr>
<tr>
<td>Ruby Mountains core complex</td>
<td>Ruby Mountains, Nevada</td>
<td>SEC 4-2-3</td>
<td>Elko, Nevada</td>
<td>−15.6</td>
<td>−115</td>
<td>Friedman et al. (2002)</td>
</tr>
<tr>
<td>Ruby Mountains core complex</td>
<td>Ruby Mountains, Nevada</td>
<td>SEC 4-2-3</td>
<td>Elko, Nevada</td>
<td>−15.6</td>
<td>−115</td>
<td>Friedman et al. (2002)</td>
</tr>
</tbody>
</table>

Note: Data compiled from Friedman et al. (1992, 2002) and Lechler and Neimi (2012). (M) in sample name represents 0.2–0.05 μm size fraction; BAD-1 (G) is a <0.2 μm size fraction. LANF—low-angle normal fault.

Figure 7. Plots showing independence of measured isotopic values in gouges and present-day meteoric water signatures. (A)–(D) $\delta^{18}O/\delta^2H$ plots for individual illitic samples. Temperatures are lower and upper temperatures used for exchanging fluid-composition calculations based on clay mineralogy. (E) $\delta^{18}O/\delta^2H$ plot for smectitic samples. (F)–(I) $^{39}$Ar/$^{39}$Ar spectra from splits of material used for isotopic analysis. Diamonds—measured isotopic values; brown lines—calculated equivalent fluid composition; blue squares—average isotopic signature of present-day precipitation (data from Friedman et al., 1992, 2002; Lechler and Neimi, 2012; compiled in Table 3). Please note: (F) in sample name indicates size fraction <0.05 μm (Stokes equivalent); (M) represents 0.2–0.05 μm size fraction.
lateral infiltration along major detachments or burial of pre-metamorphic fluids (e.g., Clark et al., 2006; Raimondo et al., 2011, 2013), we conclude that fluid circulation of crustal-scale LANF systems occurs by drawdown of meteoric waters through evolving fault and fracture networks that form and propagate in response to regional extension in the hanging wall, possibly aided by topography to drive fluid flow (Fig. 8).

Our dynamic scenario explains the observations of near-pristine meteoric water at upper- to mid-crustal levels in LANFs, with transient fault networks providing efficient pathways for significant quantities of meteoric water to reach into the crust. Our interpretation of a surface-to-depth plumbing system in LANFs and comparison with depth-to-surface fluids in thrust systems suggests that fluid dynamics of the upper crust is closely linked to the kinematic environment.

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