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Constraints on paleofluid sources using the clumped-isotope thermometry of carbonate veins from the SAFOD (San Andreas Fault Observatory at Depth) borehole

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ABSTRACT

The San Andreas Fault Observatory at Depth (SAFOD), near Parkfield, California, is a borehole drilled through two active deforming zones of the San Andreas fault, the Southwest Deforming Zone (SDZ) and the Central Deforming Zone (CDZ). These zones accommodate displacement by seismic slip and aseismic creep. Elevated fluid pressures and fluid–rock interactions have been proposed to explain the low apparent strength and aseismic creep observed, but the origin of the fluids and existence of high fluid pressures remains uncertain. We use clumped-isotope thermometry and $\delta^{18}O$–$\delta^{13}C$ compositions of calcite in veins to constrain the origin of paleofluids and compare these results to the isotopic composition of modern-day pore fluids from the SAFOD borehole and nearby areas. We observe that: (1) calcite vein temperatures vary from 81 to 134 °C, which overlaps the current ambient borehole temperatures of 110–115 °C at sampled depths; (2) vein calcite is not in carbon isotope equilibrium with modern-day pore fluids; (3) the $\delta^{18}O$ values of paleofluids close to the SDZ and CDZ, calculated from vein $\delta^{18}O$ and temperature data, are not in equilibrium with local modern-day pore waters but approach equilibrium with modern pore waters far from these zones; and (4) syntectonic vein calcite is only in C- and O-isotopic equilibrium with their host rocks within the SDZ and CDZ. Spatial patterns of $\delta^{18}O$ and $\delta^{13}C$ show little evidence for across-fault fluid-flow. Clumped isotope temperatures are consistent with locally-derived fluid sources, but not with continuous or episodic replenishment of fluids from shallow sedimentary brines or deep fluid sources. Our findings are compatible with flow of meteoric fluids from the southwestern damage zone into the SDZ and CDZ, which would have favored the formation of weak phyllosilicates and contributed to the present day weakness of the two actively deforming zones.

1. Introduction

The San Andreas Fault zone (SAFZ) accommodates displacement between the Pacific and North American Plates without producing an observable heat anomaly (e.g., Lachenbruch and Sass, 1980; Fulton et al., 2004) and under stress conditions considered unfavorable for shear failure (Zoback et al., 1987; Hickman and Zoback, 2004). These observations and others are consistent with the San Andreas Fault being mechanically weak. Numerous models have been proposed to explain this apparent weakness (Byerlee, 1990; Sibson, 1992; Sleep and Blanpied, 1992; Rice, 1992; Johnson and Mc Evilly, 1995; Kennedy et al., 1997; Hardebeck and Hauksson, 1999; Moore and Rymer, 2007; Fulton and Saffer, 2008; Noda et al., 2009; Mittempergher et al., 2011; Richard et al., 2014).

Many explanations of weakening require that fluids are present within the SAFZ, and it is well known that fluids play a critical role in controlling the chemical and physical conditions in and around fault zones (e.g., Wintsch et al., 1995; Evans and Chester, 1995). For example, high fluid pressures can reduce the effective normal stress within a fault zone causing a decrease in fault strength (e.g., Rice, 1992; Byerlee, 1993). Fluids can act to enhance mass-transfer processes within fault zones that can lead to fault weakening or strengthening (e.g., Rutter, 1976; Gratier et al., 2011). A number of studies provide evidence that fluid–rock reactions lead to the crystallization of phyllosilicates that cannot support high shear stresses due to their low coefficients of friction (e.g., Wintsch et al., 1995; Vrolijk and van der Pluijm, 1999;

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Imber et al., 2008; Lockner et al., 2011; French et al., 2015). It is likely that a combination of these processes operate within the SAFZ (e.g., Holdsworth et al., 2011; Bradbury et al., 2011; Hadizadeh et al., 2012; Richard et al., 2014; Bradbury et al., 2015).

The spatial distribution and geochemical variation of carbonate veins in the SAFZ provide constraints needed to test models of fault weakening. Stable-isotope geochemistry and fluid-inclusion analyses of syn- and post-kinematic veins are routinely used in paleofluid studies to constrain the thermal conditions during vein formation, fluid origin, and changing fluid chemistry in a number of tectonic settings (e.g. Kirschner et al., 1999; Kirschner and Kennedy, 2001; Pili et al., 2002; Boles et al., 2004; Beaudoin et al., 2011; Pili et al., 2011; Evans et al., 2012; Swanson et al., 2012; Bergman et al., 2013; Huntington and Lechler, 2015). Veins often form in fracture networks that can greatly influence permeability and fluid migration, particularly if incomplete cementation occurs (Marrett and Laubach, 1997; Laubach and Diaz-Tushman, 2009). Further, the spatial distribution of veins can be used to construct a model of the temporal variations in fluid-flow processes that could significantly influence seismic processes at depth.

The goal of this study is to determine the sources and temperatures of paleofluids that contributed to vein formation in two actively deforming zones within the San Andreas Fault (SAF) at the San Andreas Fault Observatory at Depth (SAFOD). We use conventional oxygen and carbon isotopic analyses and recently developed techniques of clumped-isotope analysis. The results provide constraints on fluid-flow models in the San Andreas Fault at seismogenic depths.

2. Background

2.1. Geologic setting of SAFOD

The SAFOD Observatory, a part of the Earthscope initiative, has provided access to rocks from the top of the seismogenic zone of the SAFZ. The SAFZ consists of a broad northwest-trending zone of deformation that accommodates right-lateral motion between the North American and Pacific tectonic plates (e.g., Titus et al., 2011). The SAFOD borehole, spudded approximately 1.8 km southwest of the surface trace of the SAFZ near the city of Parkfield, CA (Fig. 1), is located along a segment of the fault zone that exhibits both creep and microseismic-slip behavior. A vertical pilot hole was drilled to 2.2 km true vertical depth (TVD) in 2002. From the same drill pad, the main hole was drilled in three phases between 2004 and 2007. Phase 1 drilling in 2004 reached a TVD of 1.5 km along a vertical path, and then was deviated 55° to an inclined orientation of 35° towards N35°E for approximately 2.5 km TVD. Phase 2 drilling in 2005 extended the main borehole to a TVD of ~3.2 km across the geologic boundary between the Pacific and North American Plates. Phase 3 drilling in 2007 resulted in several multilaterals being drilled through the steel casing of the main borehole at ~2.5 km depth.

During Phase 2 drilling, a 200-meter wide damage zone was defined on the basis of P-wave (Vp) and S-wave (Vs) velocities and relatively low resistivity values. This damage zone is between 3157 and 3400 m measured depth (m MD; Fig. 1), within the Great Valley Group. Casing

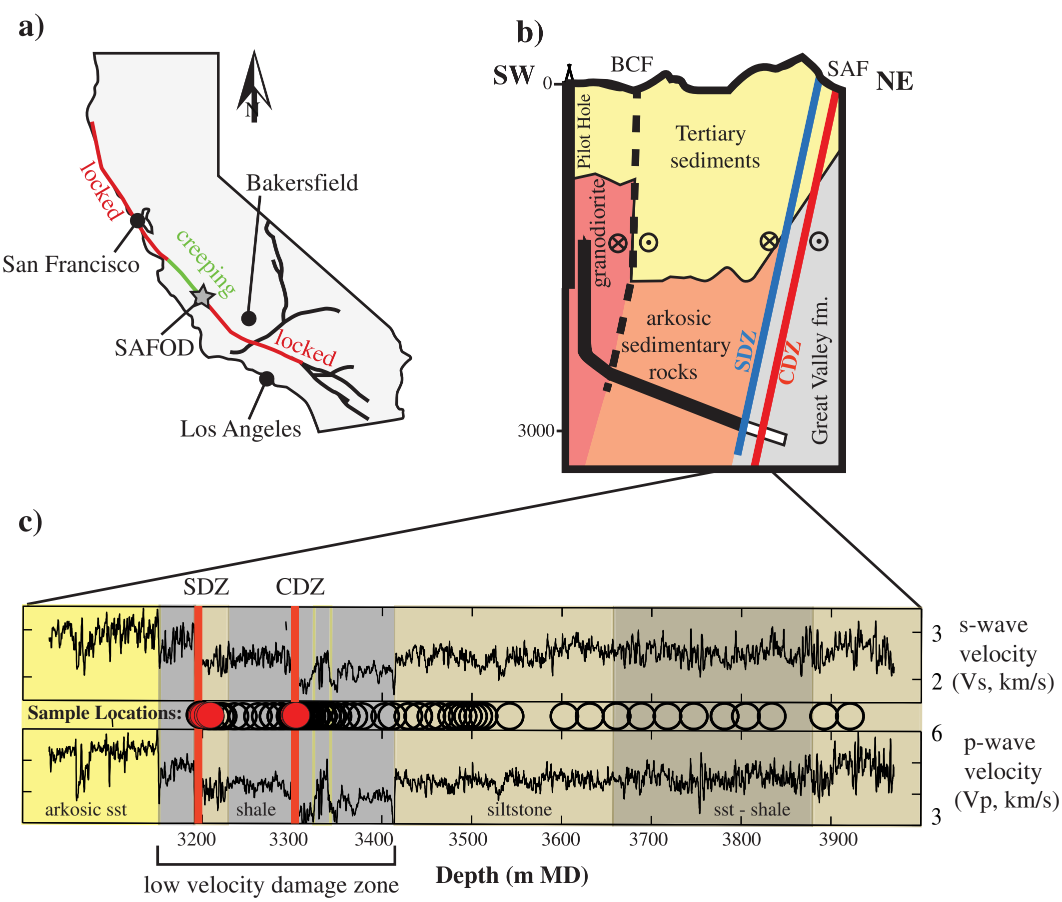

Fig. 1. Geologic setting of the SAFOD drilling site and the studied deforming zones. (a) The SAFOD borehole is located just north of Parkfield, California, at the transition between a creeping segment of the fault to the north and a locked segment of the fault to the south. (b) Generalized cross-section with the borehole geometry (modified from Bradbury et al., 2011 and Mittempergher et al., 2011). Major faults intersected include the Buzzard canyon fault (BCF) and the two active traces of the SAF. Circles containing an X indicates plate motion away from the reader and circles containing a dot indicates plate motion towards the reader. (c) Downhole geophysical logs showing the low velocity damage zone containing both the "Southern Deforming Zone" (SDZ) and "Central Deforming Zone" (CDZ) (Zoback et al., 2010). Samples in this study come from both 95 Phase II cuttings (open circles) and Phase III core (filled circles) from about 3186 to 3945 m measured depth (MD). The depths shown here, and on all other figures, are for phase 3 drilling and do not agree with phase 2 drilling depths. Phase 2 drilling depths can be calculated by (1) subtracting 5.03 m from runs 1–3, and (2) adding 3.96 m to runs 4–6 (Zoback et al., 2010).