February 2020

Shallow Water Seafloor Geodesy: GPS on An Anchored Spar Buoy

Surui Xie
University of South Florida

Follow this and additional works at: https://digitalcommons.usf.edu/etd

Part of the Geology Commons, Geophysics and Seismology Commons, and the Ocean Engineering Commons

Scholar Commons Citation

This Dissertation is brought to you for free and open access by the USF Graduate Theses and Dissertations at Digital Commons @ University of South Florida. It has been accepted for inclusion in USF Tampa Graduate Theses and Dissertations by an authorized administrator of Digital Commons @ University of South Florida. For more information, please contact digitalcommons@usf.edu.
Shallow Water Seafloor Geodesy: GPS on An Anchored Spar Buoy

by

Surui Xie

A dissertation submitted in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy
School of Geosciences
College of Arts and Sciences
University of South Florida

Co-Major Professor: Timothy H. Dixon, Ph.D.
Co-Major Professor: Rocco Malservisi, Ph.D.
Paul H. Wetmore, Ph.D.
Chad Lembke, M.M.E.
Robert C. Tyce, Ph.D.

Date of Approval:
March 3, 2020

Keywords: Geodetic monitoring, oceanographic noise, subduction zone, coastal subsidence

Copyright © 2020, Surui Xie
Acknowledgments

I could not complete this dissertation without the generous advice and support of many people. Most notably my advisor, Tim Dixon. Tim guided me in developing visions of becoming a scientist and educator. He can always offer competent advice, not only in research, but also in life. I owe much of my colorful and cheerful PhD life to Tim’s enthusiastic and optimistic spirit.

My co-advisor, Rocco Malservisi, taught me to do research with cautious attitude. There were numerous times that I was paranoid in my narrow minded understanding but he patiently explained to me with his thorough consideration. From him I learnt the importance of rigorous tests in research.

I thank my committee, Paul Wetmore, who led me several wonderful trips to the southwestern US. He is so knowledgable about this region that I feel like visiting his backyard while in the field. Interactions with Paul greatly broadened my view to the geology discipline. Chad Lembke and Robert Tyce are experts in ocean engineering, they provided great advice to my dissertation project. I am forever grateful for their time and effort.

I acknowledge my collaborators and friends. Denis Voytenko is exceptionally helpful that I can always count on him. I see him as the “Monkey King” of the Geodesy Lab. David Holland invited me to participate in Greenland field work and helped me interpret the data. Denise Holland provided meticulous logistic assistance. Lis Gallant is a cool and considerable fellow. Lewis Owen is a great scientist who worked in the field with me for several times and hosted me in his lab. Paula Figueiredo made the field work interesting. Jay Law did the tough work of assembling electronics during the deployment of the buoy. Randy Russal worked tirelessly to ensure I have sufficient data for my dissertation.
I would like to acknowledge the colleagues and staff in our department: Chuck Connor, Laura Connor, Mel Rodgers, Jochen Braunmiller, Steve McNutt, Glen Thompson, Sylvain Charbonnier, Aurélie Germa, Sarah Kruse, Mark Rains, Jun Chen, Judy McIlrath, Mandy Stuck, Jessica Wilson, Davina St. Catherine. I also want to thank Jackie Dixon and David Naar in USF College of Marine Science for their support. Thank you to the graduate students in Geophysics: Jacob Richardson, Qian Yang, Makan Karegar, Nick Voss, Anita Marshall, Christine Downs, Cassandra Smith, Alex Farrell, Heather McFarlin, Sajad Jazayeri, Kathryn Dorn, Daniel Graybeal, Amy Nachbor, Mitch Hastings, Mahsa Afra, Taha Sadeghi Chorsi, Robert Van Alphen.

Thank you to my family. Especially my wife Fanghui Deng, who supports me all the time.

This dissertation was supported by USF Teaching and Research Assistantships, and the National Science Foundation (grant 1538179 to Tim Dixon).
# Table of Contents

List of Tables iii

List of Figures iv

Abstract vi

1 Introduction 1
   1.1 Geodesy at the land-ocean margin 1
   1.2 Motivation for shallow water seafloor geodesy 2
   1.3 Previous work 4
   1.4 Outline of the dissertation 4

2 Instrumentation and Test in Tampa Bay 6
   2.1 System design 6
   2.2 Test site 8
   2.3 Deployment 11
   2.4 Data analysis 13
      2.4.1 Three-dimensional transformation to estimate anchor position 13
      2.4.2 GPS data processing 15
      2.4.3 Magnetic correction for digital compass 17
      2.4.4 Seafloor marker positioning 24
   2.5 Discussion 28
      2.5.1 Error analysis 28
      2.5.2 Response to environmental forcing 30

3 Continued Development and Potential Applications 35
   3.1 Continued development 35
      3.1.1 Motivation 35
      3.1.2 Proposed design of a system capable for deeper water 37
   3.2 Potential applications in shallow water seafloor geodesy 39
      3.2.1 Monitoring natural and human-induced offshore subsidence 39
      3.2.2 Monitoring submerged volcanoes 40
      3.2.3 Monitoring subduction offshore strain processes 41

4 Conclusion 44

References 45
Appendices  

<table>
<thead>
<tr>
<th>Appendix</th>
<th>Licence and reprint of Xie et al., 2016, JGlaciol</th>
<th>55</th>
</tr>
</thead>
<tbody>
<tr>
<td>Appendix II</td>
<td>Licence and reprint of Xie et al., 2018, TC</td>
<td>56</td>
</tr>
<tr>
<td>Appendix III</td>
<td>Licence and reprint of Xie et al., 2019, NC</td>
<td>66</td>
</tr>
<tr>
<td>Appendix IV</td>
<td>Licence and reprint of Xie et al., 2019, IGR</td>
<td>82</td>
</tr>
<tr>
<td>Appendix V</td>
<td>Copyright permission for Xie et al., 2019, JGR</td>
<td>99</td>
</tr>
<tr>
<td>Appendix VI</td>
<td>Error propogations</td>
<td>139</td>
</tr>
<tr>
<td></td>
<td></td>
<td>142</td>
</tr>
</tbody>
</table>
List of Tables

Table 2.1 Modeled horizontal displacements during three discrete events 28
## List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Figure 1.1</td>
<td>$M_w \geq 7.5$ earthquakes occurred between 1 January 1980 and 1 March 2019 in several subduction zones.</td>
<td>3</td>
</tr>
<tr>
<td>Figure 2.1</td>
<td>System design and test site</td>
<td>7</td>
</tr>
<tr>
<td>Figure 2.2</td>
<td>Bed characteristics of the test area</td>
<td>10</td>
</tr>
<tr>
<td>Figure 2.3</td>
<td>Deployment of assembled mooring</td>
<td>11</td>
</tr>
<tr>
<td>Figure 2.4</td>
<td>Above-waterline section of the GPS-spar buoy system</td>
<td>12</td>
</tr>
<tr>
<td>Figure 2.5</td>
<td>Observation geometry of the system</td>
<td>14</td>
</tr>
<tr>
<td>Figure 2.6</td>
<td>15 days of GPS and selected oceanographic data</td>
<td>16</td>
</tr>
<tr>
<td>Figure 2.7</td>
<td>Heading/pitch/roll measurements over the sampled time period</td>
<td>18</td>
</tr>
<tr>
<td>Figure 2.8</td>
<td>Heading offset and anchor position estimates for 22 September 2018</td>
<td>20</td>
</tr>
<tr>
<td>Figure 2.9</td>
<td>GPS and estimated anchor positions for a single day</td>
<td>22</td>
</tr>
<tr>
<td>Figure 2.10</td>
<td>Heading offset variations and an example of inverse perspective transformation for buoy orientation estimates</td>
<td>23</td>
</tr>
<tr>
<td>Figure 2.11</td>
<td>Top of buoy position, anchor position, and selected environmental data</td>
<td>26</td>
</tr>
<tr>
<td>Figure 2.12</td>
<td>Anchor displacement estimates</td>
<td>27</td>
</tr>
<tr>
<td>Figure 2.13</td>
<td>Anchor displacement and environmental variables during Hurricane Michael and an unnamed storm in December 2018</td>
<td>31</td>
</tr>
<tr>
<td>Figure 2.14</td>
<td>Top of buoy position, anchor position, and selected environmental data</td>
<td>33</td>
</tr>
<tr>
<td>Figure 3.1</td>
<td>Bathymetry of subduction zones in Cascadia and Central America</td>
<td>36</td>
</tr>
<tr>
<td>Figure 3.2</td>
<td>Sketch of proposed system with a mooring line</td>
<td>37</td>
</tr>
<tr>
<td>Figure 3.3</td>
<td>Change in effective length of a mooring cable</td>
<td>38</td>
</tr>
<tr>
<td>Figure 3.4</td>
<td>Bathymetry and vertical land motion of the Gulf of Mexico</td>
<td>40</td>
</tr>
</tbody>
</table>
Figure 3.5  Bathymetry and radar images of the Anak Krakatau volcano

Figure 3.6 Potential application area in Central America

Figure A1 Inverse perspective transformation of photographic images taken on 23 August 2018

Figure A2 Inverse perspective transformation of photographic images taken on 9 September 2018

Figure A3 Inverse perspective transformation of photographic images taken on 5 April 2019

Figure A4 Anchor displacement and environmental variables during periods of relatively large motion in May and June 2019

Figure A5 GPS position time series of a land site for comparison

Figure A6 Illustration of error in vertical component of anchor position estimates induced by error in GPS to anchor length measurement

Figure A7 Water temperature measured near the surface

Figure A8 Anchor displacement and current velocities along the long-term motion direction during Hurricane Michael and the unnamed storm in December 2018

Figure A9 Simulation of effect of extreme weather condition at seafloor ballast
Abstract

Measuring seafloor motion in shallow coastal water is challenging due to strong and highly variable oceanographic effects. Such measurements are potentially useful for monitoring near-shore coastal subsidence, subsidence due to petroleum withdrawal, strain accumulation/release processes in marine shelves and submerged volcanoes, and certain fresh water applications, such as volcano deformation in caldera-hosted lakes. I participated in a project to develop a seafloor geodetic system for this environment based on an anchored spar buoy topped by high precision GPS. Orientation of the buoy is measured using a digital compass that provides heading, pitch, and roll information. The combined orientation and GPS tracking data are used to recover the three-dimensional position of the seafloor marker (anchor). A test system has been deployed in Tampa Bay, Florida, for over one year, and has weathered several major storms without incident. Even in the presence of strong tidal currents which can deflect the buoy several meters from vertical, daily repeatability in the corrected three-component position estimates is 1–2 cm or better. Except for the rapid motion during the first month after deployment due to settling, other large anchor displacements correspond to extreme weather events, and are likely associated with current-induced scour activity.
1. Introduction

1.1 Geodesy at the land-ocean margin

Geodesy is a fundamental technique for measuring changes in the surface of the Earth and other planets, and their gravitational fields. Some of Earth’s fastest changes occur at or near the land-ocean margin, such as ice motion around Greenland and Antarctica, earthquakes within active plate boundaries, and subduction zone volcanic eruptions. Observing changes at these places has deepened our understanding of natural processes, and human impacts (e.g., Lawson and Reid, 1908; Oppenheimer, 2003; Stocker et al., 2013; Doocy et al., 2013).

I use space, terrestrial and marine geodetic techniques to investigate Earth’s topography and its deformation over time in critical zones, primarily applied to the study of Greenland glacier dynamics and strain processes near active plate boundaries. The geodetic techniques, including Global Navigation Satellite Systems (GNSS), Terrestrial Radar Interferometry (TRI), Light Detection and Ranging (LiDAR), and Photogrammetry, allow changes over a wide range of time scales to be quantified. Appendices I–IV list several examples of geodetic applications I have conducted, ranging from measuring glacier dynamics over minutes to fault slip rate over hundreds of thousands of years (Xie et al., 2016, 2018, 2019a,b).

As one of the most vibrant fields in Earth science, Geodesy has been evolving rapidly since the beginning of the satellite era. Advances in theory and engineering are changing the future of Geodesy. The geodetic community’s efforts to provide improved understanding of different phenomena in Earth science, such as earthquakes, magmatic activity and sea level change are leading to better forecasting of hazards (Aster et al., 2015). As a member of this
community, I have focused my dissertation project towards developing geodetic infrastructure and data processing methods that may help to address some of the grand challenges in Geodesy.

1.2 Motivation for shallow water seafloor geodesy

Space geodetic techniques such as GNSS have achieved centimeter to millimeter precision and are now widely used to study Earth surface deformation (Dixon, 1991; Segall and Davis, 1997; Reilinger et al., 2006; Tregoning et al., 2009; Li et al., 2013; Bock and Melgar, 2016; Herring et al., 2016). However, precise application of these techniques is limited to land, whereas 71% of the Earth’s surface, and most plate boundaries, are covered by water. A number of devastating earthquakes in the last a few decades occurred over the poorly monitored offshore regions (Figure 1.1). There is a clear need for high precision geodetic techniques that can work in the submarine environment (e.g., Newman, 2011; Bürgmann and Chadwell, 2014).

While several techniques for seafloor geodesy are available, they typically work best in the deeper ocean (>1 km water depth) where noise introduced by oceanographic effects is relatively low. The coastal ocean is a more challenging environment for measuring seafloor displacements, because its spatially and temporally variable oceanographic effects can be significant. Potential applications of a shallow seafloor geodetic system include monitoring of:

1) Volcano deformation: Some active volcanoes have monitoring areas partly covered by shallow water. Marine examples include Anak Krakatoa in Indonesia, Santorini caldera in Greece and Campi Flegrei adjacent to the Bay of Naples, Italy. Fresh water examples include Yellowstone Lake in USA, Lake Taupo in New Zealand, and Lago Nicaragua, which includes Ometepe Island and an active volcano, Concepción.

2) Offshore oil fields: Oil and gas field management can lead to significant uplift and/or subsidence. Accurate monitoring can help to assess reservoir performance and infrastructure integrity.
Figure 1.1: $M_w \geq 7.5$ earthquakes occurred between 1 January 1980 and 1 March 2019 in several subduction zones, shown with beach balls. (a) Aleutian. (b) East Japan. (c) Indonesia. (d) New Zealand. (e) Central America. (f) Cascade. Earthquake data were downloaded from USGS.
3) Strain accumulation and release processes in shallow marine shelves associated with seismic hazard: Many shallow marine shelves could host earthquakes. Shallow seafloor geodetic measuring system can potentially improve seismic hazard assessments.

1.3 Previous work

Several methods have been developed to measure seafloor motion, including:

1. Bottom pressure (e.g., Chadwick Jr et al., 2006; Chierici et al., 2016);
2. GPS-acoustic (e.g., Spiess, 1985; Spiess et al., 1998; Chadwell and Spiess, 2008; Sato et al., 2011; Yokota et al., 2016);
3. Direct-path acoustic ranging (e.g., Chadwick Jr and Stapp, 2002; Osada et al., 2012; McGuire and Collins, 2013);
4. Multibeam sonar surveying (e.g., DeSanto et al., 2016; Fujiwara et al., 2017);
5. Strainmeter and tiltmeter systems (e.g., Anderson et al., 1997; Zumberge et al., 2018).

While all of these techniques are applicable in certain situations, they would have limitations in shallow, turbulent coastal water due to oceanographic noise. To complement existing methods, Istituto Nazionale di Geofisica e Vulcanologia (INGV, the Italian Institute of Geophysics and Volcanology) has developed a system of GPS-tracked rigid buoy capable of measuring vertical seafloor displacement in water depths shallower than 150 meters (De Martino et al., 2014; Iannaccone et al., 2018). Currently they operate four such buoys close to Campi Flegrei, Italy. They use weekly averaged values of the vertical displacement component to estimate seafloor uplift associated with volcanic deformation, achieving performance that is comparable to on-land GPS.

1.4 Outline of the dissertation

We have modified the INGV design in order to determine both vertical and horizontal components of displacement and to reduce cost of the instrument. In this dissertation, I describe the design, construction, and initial results of the system, assess some potential
applications, and outline future developments. The first chapter introduces the background. In the second chapter, I report the instrumentation and system test in Tampa Bay, Florida. In chapter three, I propose continued development of the system, and several potential applications in shallow water seafloor geodesy. The last chapter summarizes the work we have done. A concise version describing the research has been published in the *Journal of Geophysical Research: Solid Earth* as: Xie, S., Law, J., Russell, R., Dixon, T.H., Lembke, C., Malservisi, R., Rodgers, M., Iannaccone, G., Guardato, S., Naar, D.F., Calore, D., Fraticelli, N, Brizzolara, J., Gray, J.W., Hommeyer, M., Chen, J. (2019), Seafloor geodesy in shallow water with GPS on an anchored spar buoy, J. Geophys. Res. Solid Earth, 124, 12116-12140, doi: 10.1029/2019JB018242. License is provided in Appendix V.
2. Instrumentation and Test in Tampa Bay

2.1 System design

Our system design is based on the successful MEDUSA research and monitoring marine infrastructure (Multiparametric Elastic-beacon-based Devices and Underwater Sensor Acquisition system) used by INGV to monitor vertical deformation of the seafloor in the Gulf of Pozzuoli in southern Italy, adjacent to the Campi Flegrei volcanic area (De Martino et al., 2014; Iannaccone et al., 2018). Their infrastructure uses a GNSS receiver installed on top of each buoy, connected to the seafloor by either a rigid spar, or in deeper water (>40 m), a rigid spar plus a steel cable. Assuming that buoyancy maintains a near-vertical orientation, vertical displacement of the seafloor can be estimated by correcting for the fraction of measured surface vertical motion induced by surface horizontal motion of the buoy using simple geometry (De Martino et al., 2014). We add heading/pitch/roll measurements to the system and perform a 3-dimensional transformation to recover both horizontal and vertical components of the seafloor anchor motion (ballast). Our system is called SUBGEO (Shallow Underwater Buoy for Geodesy) and can be used in coastal regions shallower than 40 m. It can be adapted to work in deeper water (<200 m) by adding a cable between the buoy and the ballast, and performing additional orientation measurements.

The use of high precision GPS systems for measuring seafloor movement requires both stability and a direct connection to the seafloor, while also keeping the antenna and electronics safely above the waterline in all anticipated sea states. These requirements necessitate a design with a long spar connecting the GPS antenna to the seafloor, excessive net flotation for stability, and significant ballast to hold the system in place. Figure 2.1a shows the system design. The above-waterline section of the buoy consists of a superstructure (upper part)
Figure 2.1: System design and test site. (a) System design. GPS is on top of the buoy, and associated electronics are mounted in a box below. The system is powered by solar-charged batteries. A float provides buoyancy, keeping the buoy close to vertical. The buoy is attached to a seafloor ballast through a large shackle. (b) Bathymetry of Tampa Bay, Florida. Markers show the test site and other nearby instruments. Note location of the GPS-buoy close to a tidal outflow channel.

and access ladder. A GPS antenna is mounted on top at the center of the superstructure. A GPS receiver, a digital compass (Honeywell HMR3000) and associated electronics are installed in a fixed weather-proof enclosure rigidly connected to the superstructure, adjacent to exterior-mounted solar panels. This section of the buoy is designed to be lightweight and have a low cross-section area to minimize wind stress. However, there is sufficient space in the superstructure to incorporate additional instruments if required. Below the waterline, a polyurethane foam float is installed at a depth that ensures that it is continuously submerged regardless of sea state. Combined with increased displacement achieved by welding caps on the individual spar components, a net buoyancy of 9 tons is provided to maintain the buoy in a nearly vertical position. At the bottom of the system, a $2.4 \text{ m} \times 2.4 \text{ m} \times 1.7 \text{ m}$ concrete
ballast reinforced by steel rebar is used as an anchor, attached to the buoy with a shackle. This concrete provides a ballast of \( \sim 13 \) tons when submerged in seawater. Therefore, when the buoyancy and ballast are combined, a net ballast (negative buoyancy) of more than 4 tons is achieved.

2.2 Test site

The test site is located on the bay side of Egmont Key within Tampa Bay, Florida at a water depth of \( \sim 23 \) m (Figure 2.1b). This section of the West Florida continental shelf is part of a broad, mostly submerged ancient carbonate platform with a thin layer of unconsolidated sediment over Miocene limestone (Doyle and Sparks, 1980; Evans et al., 1985; Hine, 1997, 2013; Hine et al., 2008; Berman et al., 2005). While the site is protected from some of the wave energy from the Gulf of Mexico by Egmont Key, it experiences strong ebb and flood tidal currents (Berman et al., 2005, and references therein). Maximum current speeds in the channels adjacent to Egmont Key range from 1.8 m/s on the ebb tide to nearly 1.1 m/s on the flood tide (Stott and Davis Jr, 2003). These currents can cause correspondingly large deflections (>3 m) of the buoy from the vertical position and hence provide a rigorous test of our approach to horizontal motion corrections. The site is adjacent to the busy Egmont shipping channel (Berman et al., 2005; Gray, 2018) and occasionally experiences wake from passing ships.

The GPS-buoy deployment location is at the southern edge of a \( \sim 30 \) m deep known as Egmont Deep (previously Egmont Hole; Figure 2.1b). This Deep and its connection to the Egmont Channel is analogous to the geologic term “gorge” or “throat”, which is a deep scoured bathymetric feature found at most barrier island inlets (or passes) (Berman et al., 2005, and references therein). The deepest part, the gorge, is formed and maintained by the strong tidal currents entering and exiting Tampa Bay. The formation of Egmont Deep is likely the result of a combination of karst and scour processes (Berman et al., 2005). The limestone bedrock here was subaerially exposed repeatedly during previous sea level low stands. Following the most recent ice age, sea level rose and flooded Tampa Bay, with
subsequent sediment infilling of the deeps and channel scouring. Sediments adjacent to Egmont Deep consist primarily of coarse carbonate shell fragments, and siliceous sand and silt (Ginsburg and James, 1974; Doyle and Sparks, 1980; Brooks and Doyle, 1998; Berman et al., 2005).

SCUBA observations made in December 2000 showed a seafloor nearly devoid of sediment, composed of smooth limestone with phosphatic nodules and small scour pits (Berman et al., 2005). Only large, rubble-sized limestone debris, oyster shells, and human trash were observed as unconsolidated materials in or near the deepest areas. Further north, a 30-m transect showed multi-colored sponges. In April 2001, another SCUBA survey (while retrieving a deployed ADCP) showed a surface buried by ~0.5 m of shell fragments and coarse sediment, with previous rubble-sized debris no longer visible, possibly buried (Berman et al., 2005). These observations suggest a highly dynamic seafloor environment, presumably due to strong tidal currents, with any fine-grained sediment entering Egmont Deep periodically transported out of the area. The buoy location is immediately southeast of the deepest, high velocity part of the channel, and at the present time contains some reworked sediments, subaqueous dunes, and smaller sedimentary bedforms (Figure 2.2a). Diver observations on October 1, 2019 showed the presence of unconsolidated sandy sediment at least 1.5 meters thick, based on the length of rebar driven into the sediment.

Berman et al. (2005) compare bathymetry collected using the same multibeam system (Kongsberg Simrad EM3000 300 kHz system) from surveys made in 1999 and 2001. They observed large dunes near the study site with 2 m vertical relief, an average slope of 6°, and average wavelength of 150 m, with smaller superimposed dunes with typical relief of 0.3 m, wavelength of 5-9 m, and length of ~50–150 m. At the study area, smaller dunes are observed with some slopes up to ~10°. In April 2018, multibeam data from a Reson SeaBat T50-R dual-head 200–400 kHz system (run at 400 kHz) were collected in the study area. These data also show the presence of numerous small subaqueous dunes near the buoy location (Figure 2.2b).
Figure 2.2: Bed characteristics of the test area. (a) Bed aspect map derived from multibeam sonar survey data. Aspect changes from light blue to red indicate subaqueous dunes. Aspect of $0^\circ$ and $360^\circ$ denote bed facing north. (b) Image of the seafloor at the test site during deployment. Diver’s hand (in black neoprene glove) is penetrating several centimeters of soft sediment. Sediment here is primarily sand, with a thin muddy layer and some organisms above it. The unconsolidated sediment is at least 1.5 meters thick.
2.3 Deployment

As detailed in the section 2.1, the system’s high-resolution GPS application demands a moored structure with net flotation and ballast mass to provide sufficient stability while also keeping the antenna and electronics above the water line in all sea states. The system’s spar length of \(\sim 30\) meters and combined in-air weight of \(\sim 27\) tons makes the deployment logistically complicated. For this reason Orion Marine Group of Tampa was contracted to perform the deployment using a 46-m barge equipped with a 250-ton crane.

While the system is designed to be deployable in two stages, allowing a fully instrumented tower to be bolted onto the uppermost spar flange after it is in place, an alternate plan was used for our deployment. Strong tidal currents at the mouth of Tampa Bay introduced
potential issues with aligning the two flanges near the waterline in a two-stage deployment. For this reason, the system was deployed fully assembled except for the electronics. The electronics were then installed via small boat after the buoy was in place.

The use of a large crane with an auxiliary line makes a single-stage deployment possible. As shown in Figure 2.3, the anchor is lifted via the main winch line while an auxiliary line supports the spar section at two lifting points with a custom sling. This method minimizes the time required for the crane and barge to be on site. The entire deployment including a 3-hour transit was accomplished in approximately 6 hours, with an additional 6 hours required for installation of power and electronics components. The deployment was executed successfully on August 23, 2018, with data transmitting 6 hours after the anchor was placed on the bottom. Figure 2.4 shows the above-waterline section of the system.

Figure 2.4: Above-waterline section of the GPS-spar buoy system.
2.4 Data analysis

GPS data are collected at 15-second intervals. Heading/pitch/roll are measured by the digital compass every 5 seconds with 0.1° resolution. All data are downloaded through a Freewave radio link ∼2.5 km away from the buoy. Here we report data obtained between 23 August 2018 and 24 August 2019 (Dixon et al., 2019). Data gaps during this period occurred <0.01% of the time.

2.4.1 Three-dimensional transformation to estimate anchor position

The ballast serves as a seafloor marker, and its displacement is used as a proxy for seafloor motion. Because of its large mass, it should be relatively stable except during extreme sea states or weather events. Figure 2.5 shows the system geometry. Anchor position \((N_a, E_a, U_a)\) can be calculated using:

\[
\begin{bmatrix}
N_a \\
E_a \\
U_a
\end{bmatrix} =
\begin{bmatrix}
N_g \\
E_g \\
U_g
\end{bmatrix} +
\begin{bmatrix}
N_{ag} \\
E_{ag} \\
-U_{ag}
\end{bmatrix}
\] (2.1)

where \((N_g, E_g, U_g)\) are north/east/up components of the GPS position, and \((N_{ag}, E_{ag}, U_{ag})\) represent anchor coordinates defined in a local Cartesian coordinate system G-X_bY_bZ_b whose origin is at the GPS antenna phase center, and whose three axes point to geographic north/east/down (shown in Figure 2.5a by blue/cyan/red colored arrows). \((N_{ag}, E_{ag}, U_{ag})\) can be calculated using a 3-axis rotation:

\[
\begin{bmatrix}
N_{ag} \\
E_{ag} \\
U_{ag}
\end{bmatrix} = R_z(-\alpha)R_y(-\beta)R_x(-\gamma)
\begin{bmatrix}
0 \\
0 \\
L
\end{bmatrix}
\] (2.2)

where \(L\) is the length of the buoy (defined as length from the GPS antenna phase center to the pivot point of the anchor). \(\alpha, \beta,\) and \(\gamma\) are measured heading, pitch, and roll angles of
Figure 2.5: Observation geometry of the system. (a) State of the system when the buoy (represented by thick gray line) is vertical and three axes of the digital compass are at initial orientations (heading/pitch/roll measurements are all zero). Signs of these measurements follow the right-hand rule. (b) State of the system when GPS and digital compass modules are perturbed from initial positions. Solid color arrows show perturbed orientations of heading/pitch/roll axes. Dashed color arrows show reference directions.

the buoy. Their signs are defined by the right-hand rule (Figure 2.5a). The three rotation matrices in Equation 2.2 are calculated by:

\[
R_x(-\gamma) = \begin{bmatrix}
1 & 0 & 0 \\
0 & \cos(\gamma) & -\sin(\gamma) \\
0 & \sin(\gamma) & \cos(\gamma)
\end{bmatrix}
\]  

(2.3)

\[
R_y(-\beta) = \begin{bmatrix}
\cos(\beta) & 0 & \sin(\beta) \\
0 & 1 & 0 \\
-\sin(\beta) & 0 & \cos(\beta)
\end{bmatrix}
\]  

(2.4)
\[ R_z(-\alpha) = \begin{bmatrix}
\cos(\alpha) & -\sin(\alpha) & 0 \\
\sin(\alpha) & \cos(\alpha) & 0 \\
0 & 0 & 1
\end{bmatrix} \quad (2.5) \]

Based on Equations 2.1 – 2.5, for a given time, if GPS position and heading/pitch/roll of the superstructure are measured, position of the anchor can be estimated. Note that heading in these equations is relative to geographic north, therefore a correction for magnetic declination is needed.

### 2.4.2 GPS data processing

Most of our GPS data are processed using the TRACK v1.30 kinematic processing software (Chen, 1998; Herring et al., 2018). We also processed some data using the free online GPS processing tool CSRS provided by the Canadian Geodetic Survey (https://webapp.geod.nrcan.gc.ca/geod/tools-ouils/ppp.php?locale=en, last access on 23 September 2019) for comparison. For TRACK processing, a stationary GPS site 35 km away from the buoy is used as reference station (red triangle in Figure 2.1b). Our kinematic site positions, representing the instantaneous position of the phase center of the GPS on top of the buoy, are estimated in long baseline mode, with motion modeled as a random walk. We set allowable changes in velocity for all three components to be 1 m per second, approximately the same as the speed of a typical tidal current in the bay. A cut-off angle of 15 degrees is used to reduce the influence of multi-path. GPS position is estimated at 15 second intervals, the sampling rate of the GPS observations.

Our GPS raw data are written as daily files. To minimize potential jumps at the boundaries of each day due to smoothing gaps, we use 30 hours of data centered on the middle of each day to form session files. For a few known jumps (caused by anchor slip immediately after deployment, hurricanes, or periods of other extreme weather), we omit data within 2 hours of the visually inspected large jumps and do not allow sessions to span the jumps. After finishing each processing session, we remove position estimates the day
Figure 2.6: 15 days of GPS and selected oceanographic data. Black dots show processed GPS positions using TRACK kinematic processing software (Chen, 1998; Herring et al., 2018). Red dots are results processed using the online tool of the Canadian Spatial Reference System (CSRS) Precise Point Positioning (PPP). Except for a few outliers, solutions from these two software programs match well. Data gaps in TRACK solutions are mainly due to data gaps in the on-land reference station. Cyan curves in the upper two panels show surface tidal current speeds from the nearby current meter shown in Figure 2.1b, blue curves show the Tampa Bay Coastal Ocean Model (TBCOM) hindcast surface current speeds at the GPS-buoy location (Chen et al., 2018, 2019). GPS motion is positively correlated with current velocity (the linear correlation coefficient between GPS displacement and modeled current speed is 0.56 for the north component and 0.80 for the east component).

before and after (data during 21:00–24:00 in the previous day and 00:00–03:00 in the next day), and epochs with any ambiguities unfixed to integer values are deleted. Typical formal error estimates for a single epoch are 1.5–2.0 cm for the horizontal components and 4–5 cm for the vertical component. Note that TRACK uses differential phase measurements to
estimate GPS positions, thus observations from both the kinematic and reference stations are required. Data gaps (mainly due to data gaps in the reference station) lead to solution gaps but only account for a small fraction of the resulting displacement time series.

Black dots in Figure 2.6 show the kinematic GPS positions processed using TRACK. Red dots show solutions by CSRS for comparison. CSRS uses a precise point positioning (PPP) strategy and does not require reference stations. Solutions from these two processing streams show good agreement (Figure 2.6), with relatively large differences mostly occurring when tidal current speed is high and tilt of the antenna is largest. These points are flagged as outliers (below) and not used for anchor position estimates. Due to the advantage of not requiring a reference station, there are fewer gaps in the CSRS solutions. However, currently the online CSRS-PPP tool needs individual data files to be uploaded manually. We therefore use the CSRS-PPP tool to fill gaps in TRACK results caused by outages at the reference station.

2.4.3 Magnetic correction for digital compass

The digital compass module consists of three magneto-resistive sensors oriented in orthogonal directions plus a 2-axis fluid tilt sensor, creating tilt-compensated heading, pitch, and roll data. The tilt sensor works best when kept near level (Honeywell International Inc., 2019). Although the digital compass can update continuous data with a frequency up to 8 Hz, fast jarring of the sensor degrades pitch/roll measurements and affects tilt compensated headings because not all the fluid can immediately return to the bottom of the tilt sensor’s glass ampoule (Honeywell International Inc., 2019). Therefore heading/pitch/roll measurements are less accurate when pitch or roll are large. Figure 2.7 shows outputs from the digital compass. During the observation period, pitch/roll measurements are all within ±20° and mostly ±5°. With this range, the nominal repeatability for a heading measurement should be ±0.3°, while pitch and roll measurements should be ±0.2° according to the manufacturer (Honeywell International Inc., 2019). We assume that within a short period (e.g., <1 minute in our case) the buoy moves smoothly, and use a 1-minute
Figure 2.7: Heading/pitch/roll measurements over the sampled time period. Grey dots are observed time series. Magnetic declination in heading measurements is corrected using the World Magnetic Model WMM2015v2 (https://www.ngdc.noaa.gov/geomag/WMM/soft.shtml#downloads, last access on 23 September 2019). Red lines show 1 minute (13 data points window) Gaussian filtered time series. See Figure 2.8b for example data in 1 day.

Gaussian filter to smooth heading/pitch/roll measurements. Red lines in Figure 2.7 show smoothed data. This smoothing significantly reduces the scatter of heading/pitch/roll observations (Figure 2.8b). We corrected magnetic declination using the World Magnetic Model WMM2015v2 (https://www.ngdc.noaa.gov/geomag/WMM/soft.shtml#downloads,
last access on 23 September 2019), which is about -5.4° during the observation period at the GPS-buoy location. Thus all heading data presented here are relative to geographic north.

While pitch/roll measurements are controlled by the force of gravity on fluid in the tilt sensor, heading measurements represent changes of buoy orientation, and are sensitive to the local magnetic environment and its changes. Therefore, variation of local magnetic environment (e.g., caused by orientation of the superstructure) can induce offset in the heading measurements. Thus, heading output from the digital compass may be offset from the true heading direction. Leaving this offset uncorrected would impart a significant systematic error to our tilt correction, affecting the precision and accuracy of the horizontal component estimates for the ballast. In order to avoid this bias we applied the following corrections.

We assume that for a short period (e.g., 1 day), offset in the heading measurement is constant, and the anchor is stationary or moves linearly. Offset of the heading measurement is estimated by minimizing the weighted one standard deviation (SD) of anchor position residuals. As mentioned above, outputs from the digital compass are less accurate when the tilt angle is large. GPS solutions are also noisy when pitch or roll is large (see red dots in Figure 2.8a, corresponding to periods of high-speed tidal currents). We therefore mask out data when either pitch or roll exceed 2° in any direction. A grid search with a step width of 0.1° (resolution of the digital compass measurements) is used to estimate the heading offset. Black dots in Figure 2.8c show anchor position time series for a typical day with estimated heading offset corrected, and show considerable improvement. Figure 2.9 shows anchor position for the same day in a plan view. For comparison, blue dots in Figure 2.8c show position time series assuming no instrumental offset for the digital compass. Note that we apply a modified Z-score method (Iglewicz and Hoaglin, 1993) to detect outliers in anchor position estimates for each session. For a given period, Z-score of the \( i \)-th anchor position estimate \( x_i \) is:

\[
Z_i = 0.6745|x_i - \bar{x}|/\text{MAD} \tag{2.6}
\]
Figure 2.8: Heading offset and anchor position estimates for 22 September 2018. (a) North/east components of GPS time series, red marks data when pitch or roll measurements are $>2^\circ$ or $<-2^\circ$. (b) Heading/pitch/roll measurements. Note due to the 0.1° resolution of digital compass, these observations have small steps changes around the Gaussian filtered time series. Large pitch/roll are observed when current speed is high. (c) Blue dots show calculated positions using heading data corrected for magnetic declination. Black dots are calculated positions by adding an offset to heading data (magnetic declination corrected) that minimizes scatter of one day anchor positions. Red corresponds to pitch or roll exceeding 2°, orange dots are detected outliers. Weighted one standard deviation (SD) for north and east components are shown. Note that different offsets in heading do not affect estimates of vertical component, thus we do not show vertical positions here. (d) Weighted one standard deviation (root of the sum square of north and east components) when adding different offsets to digital compass measurements for 22 September 2018. Inset shows for a wider range of heading offset, units of X and Y axes are the same as (d).
where $\tilde{x}$ is the median value of all position estimates during the period and MAD denotes the median absolute deviation:

$$\text{MAD} = \text{median}(|x - \tilde{x}|)$$

with $x$ denoting a list of all position estimates during the period. We set a threshold Z-score of 3.5, as recommended by Iglewicz and Hoaglin (1993). Any of the north/east/up components with Z-scores exceeding this value are considered outliers and removed. Orange dots in Figure 2.8c show detected outliers for the example day.

The pitch/roll measurements should not have significant offsets since they use a gravitational reference. We also tested possible corrections to these measurements using the same method as for heading offset, but the best estimates are mostly within $\pm0.2^\circ$ of the measured values (Figure 2.8d shows an example). Thus, we assume no offset in pitch/roll measurements in the following analysis. Possible residual errors related to measurement of heading/pitch/roll are analyzed in the discussion section.

The black line in Figure 2.10a shows the estimated heading offset (modeled as a constant offset for each session) during the entire observation period. Except for days with flagged jumps, they are shown with daily increments. Note that there are considerable changes in the first few months, some of which are caused by rapid re-orientation of the buoy as it settled to a stable position, as well as a step change associated with a major unnamed storm around Dec 20, 2018.

We tested the validity of our heading correction technique using an independent estimate based on optical image analysis. While such an analysis would not be possible in most offshore locations, it is useful here as a check on the least squares estimation procedure described above. The image analysis uses an inverse perspective transformation (done with OpenCV-Python module: https://pypi.org/project/opencv-python/, last access on 3 February 2020) of photographic images taken from a boat or an Unoccupied Aerial Vehicle that includes the buoy and known points on nearby Egmont Key (Figure 2.10b). The thin
Figure 2.9: GPS and estimated anchor positions for a single day (22 September 2018). (a) GPS position time series. (b) Estimated anchor positions. Grey box outlines an area shown in (c). (c) Black and red circles mark 68% (7.1 cm) and 95% (12.8 cm) percentages of the distances from the daily median position. Note that (a) and (b) are of the same spatial scale. (c) is a close-up of the outlined box in (b).
Figure 2.10: Heading offset variations and an example of inverse perspective transformation for buoy orientation estimates. (a) Black line show heading offset (compass forward axis azimuth direction minus compass measured heading) estimates by minimizing daily anchor displacements (see an example in Figure 2.8c), shown with daily increments. Red dots with error bars ($2\sigma$) are estimates using inverse perspective transformation method shown in (b-d) and Appendix Figures A1–A3. (b) Image taken by an Unoccupied Aerial Vehicle (UAV) on the deployment day. P1–P4 are 4 points used to compute the inverse perspective transform parameters, corresponding to reference points 1–4 in (c). These 4 points are: P1–buoy at waterline, P2–northern corner of a pier, P3–northern corner of an old pier, P4–lighthouse. Three cyan lines (parallel in 3-D space) are used to estimate a vanishing point in UAV camera perspective projection. Red line passes through P1 and the vanishing point is then parallel to the heading axis of digital compass, its projection in orthographic projection is shown in red in (d). (c) Orthographic-imagery of the test area, downloaded through the USGS EarthExplorer (https://earthexplorer.usgs.gov/, last access on 3 February 2020). Reference 1–4 correspond to P1–P4 in (b). (d) Image of (b) in orthographic projection. Appendix Figures A1–A3 show additional examples of inverse perspective transformation from photographic images to maps.
bars of the superstructure are parallel or perpendicular to the forward direction (origin direction for heading measurements, shown by dashed arrow in 2.5a) of the digital compass. Their orientations are used to construct parallel lines that intersect at a vanishing point (estimated using least squares) in a camera perspective projection (Figure 2.10b). The buoy-water surface contact and the vanishing point thus form a line (red line in Figure 2.10b) that is parallel to the forward direction of the digital compass. The orientation of its projection can be estimated using the georeferenced image (Figure 2.10 c and d). Using this method and multiple images taken during each site visit, variation of the offset in heading measurements can be estimated (Figure 2.10 and Appendix Figures A1–A3). Red dots and error bars in Figure 2.10a represent the mean and two standard derivation of estimates for each visit. This approach yields results that are consistent with the heading offset estimation procedure described above using least squares. Large changes in heading offset occurred in the first month after deployment, as well as several days of extreme weather.

2.4.4 Seafloor marker positioning

We apply the estimated heading offsets and use Equations 2.1–2.5 to recover anchor positions for the entire period. Figure 2.11a shows the full period of the observed GPS time series (top of buoy). Figure 2.11b shows time series for the corresponding anchor position, with 15-second intervals in dark grey, and daily median time series in red. We use the median position of each day as the best estimate, and the weighted one standard deviation (SD) as a measure of uncertainty (1σ) for each day. Daily solutions and uncertainties are shown in Figure 2.12a. During the first month after deployment, the system experienced rapid horizontal motion of several meters and vertical subsidence of ~0.5 meter. In the next seven months, the anchor was relatively stable except for short perturbations associated with extreme weather. Large perturbations occurred due to passage of Hurricane Michael (around October 10, 2018) or during an extreme weather event with heavy precipitation (around December 20, 2018) (Figure 2.11c). During the following summer (June, July, August 2019) there were frequent extreme weather events (see precipitation data in Figure
2.11c), causing significant anchor motion. Several other large displacements occurred during periods of exceptionally high current periods (Appendix Figure A4). In general, long-term displacement of the anchor mimics long-term displacement of the GPS (Figures 2.11 a and b), since the anchor constrains GPS position by a rigid connection.

To assess the repeatability of seafloor marker positioning using our GPS-buoy system, we analyze timeseries for the ~5 month period from 25 December 2018 to 1 May 2019 (Figure 2.12b) when no major storms affected the area. The displacements are modeled as simple functions, then the standard deviation is computed to assess repeatability. The vertical position of the anchor is modeled as exponentially decreasing subsidence. The horizontal position components of the anchor are assumed to behave linearly with near-zero velocity, except for several discrete steps. Displacements are modeled following Larson et al. (2004):

\[
x(t_i) = A + B e^{C t_i} + \sum_{k=1}^{n} \frac{D_k}{2} (\tanh \frac{t_i - T_k}{\tau_k} - 1)
\]

where \(x(t_i)\) is the anchor position estimate at time \(t_i\). \(A, B, C,\) and \(D_k\) are parameters to be estimated. \(D_k, T_k, \tau_k\) are the displacement, middle time, and duration half width of the \(k\)-th transient events respectively. Horizontal displacements occurred in three discrete events (Table 2.1). We set \(B = 0\) for horizontal components assuming settling-related displacement is trivial in horizontal displacements during the selected period, and set \(D_k = 0\) for the vertical component assuming transient event related displacement is insignificant during the selected period. We use visually inspected values for \(T_k\) and \(\tau_k\). Pink lines in Figure 2.12a mark the selected period shown in Figure 2.12b. Dashed pink lines in Figure 2.12a mark selected transient events modeled in Equation 2.8, chosen by visual inspection. Red curves in Figure 2.12b show best fitting curves to the daily time series. Weighted one standard deviation of the residual time series for each component is used as a measure of repeatability. For horizontal components, the standard deviation is about 1–2 cm, and is larger in the east-west direction, where tidal current amplitude is much larger than in the
Figure 2.11: Top of buoy position (a), anchor position (b), and selected environmental data (c). (a) Dark grey dots show GPS positions when both pitch and roll measurements are between -2° and 2°. Pink dots are when pitch or roll exceed ±2°. Red dots are daily medians from the data shown by dark grey dots. (b) Dark grey dots using data when both pitch and roll measurements are between -2° and 2°, pink dots correspond to pink dots in (a), light grey dots are detected outliers. Red are daily medians from the data shown by dark grey dots. (c) Water level and precipitation records near the test site. Blue curve shows observed water level (location of tide gauge shown in Figure 2.1b). Red curve shows 0.2 cycle-per-day low-frequency-pass filtered water level. In the lower panel, precipitation data are daily accumulated records (location of station shown in Figure 2.1b).
Figure 2.12: Anchor displacement estimates. Black dots are session medians (mostly daily except for interrupted sessions), with $1\sigma$ uncertainties shown by light grey color bars. (a) Time series of the entire analyzed period. Solid pink lines mark the period shown in (b). Dashed pink lines mark three slip events modeled in (b). (b) Time series for the period 25 December 2018 to 1 May 2019. Red curves are best fitting curves using Equation 2.8. Weighted one standard deviation (SD) of model residuals during this ~5 month period is calculated to assess repeatability.
north-south direction (Figure 2.6). For the vertical component, the standard deviation is less than 1 cm. The horizontal component repeatability of our GPS-buoy system is about 10 times larger than a nearby land site (1.4 mm), while the vertical component repeatability is at a similar level (4.8 mm) (Appendix Figure A5).

Table 2.1: Modeled horizontal displacements during three discrete events

<table>
<thead>
<tr>
<th>Time</th>
<th>North disp. (cm)</th>
<th>East disp. (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2019-02-19</td>
<td>-2.4±0.4</td>
<td>-7.2±0.6</td>
</tr>
<tr>
<td>2019-03-17</td>
<td>-3.6±0.4</td>
<td>-9.1±0.8</td>
</tr>
<tr>
<td>2019-04-10</td>
<td>-0.3±0.4</td>
<td>-8.0±0.8</td>
</tr>
</tbody>
</table>

2.5 Discussion

2.5.1 Error analysis

Error sources for the anchor position estimates consist of two types: 1) GPS measurement and processing errors. 2) 3-dimensional transformation errors due to input parameter errors, i.e., errors in \((N_{ag}, E_{ag}, U_{ag})\) using Equation 2.2. While we apply relatively loose constraints on the GPS data processing (we model GPS motion as a random walk and allow all three components to move up to 1 meter per second), solutions from TRACK and CSRS show good consistency. The formal errors in TRACK outputs are typically 1.5–2.0 cm for horizontal components and 4–5 cm for vertical components, comparable to other kinematic GPS applications (Chen, 1998; Watson et al., 2008; Davis et al., 2014). Here we analyze errors associated with 3-dimensional transformation in Equation 2.2, which are likely to be the dominant error source.

In Equation 2.2, we assumed GPS and the anchor are perfectly on a line that crosses the long axis of a rigid buoy, therefore the anchor coordinates in a reference frame defined by the GPS and three rotation axes (G-X\(_b\),Y\(_b\),Z\(_b\) in Figure 2.5) were \((0, 0, L)\). In reality, either GPS or the pivot of the shackle may be located at a slightly biased position and \(L\) contains measurement error. That is, the anchor coordinates in G-X\(_b\),Y\(_b\),Z\(_b\) are \((p, q, L)\), where \(p, q, L\) represent anchor point coordinates along the roll/pitch/heading axes of the
digital compass. Thus there are 6 parameters \((p, q, L, \alpha, \beta, \text{and } \gamma)\) on the right side of Equation 2.2. Errors in these parameters will be propagated to \((N_{ag}, E_{ag}, U_{ag})\). Assuming the 6 parameters are independent, we rewrite Equation 2.2 in Appendix Equation A1 and take the partial derivatives of \((N_{ag}, E_{ag}, U_{ag})\) in Appendix Equations A2–A4. Some of the items can be ignored in the total error budget, some will counteract each other, and some will only cause a systematic error for all position estimates that will not affect displacement estimates. We ignore terms where more than two small values multiply together. For example, \(\cos(\alpha)\sin(\beta)\sin(\gamma)\Delta p\) can be ignored because both pitch, roll, and \(\Delta p\) are small values and the total contribution of the multiple is much less than 1 mm. Due to high nonlinearity, we do not derive detailed errors using time-varying heading/pitch/roll measurements. Instead, we apply loose constraints and derive possible maximum errors in the seafloor anchor position. For example, \(\cos(\alpha)\) or \(\sin(\alpha)\) in Appendix Equations A2–A4 should fall between -1 and 1, and they can counteract each other, but we allow their absolute values to be 1 at the same time, thus the derived errors represent the maximum possible error. Appendix formulas A5–A7 show the possible amplitudes of anchor positioning errors. Among these, errors in heading measurements will not affect the vertical component of the anchor position estimates. An error in buoy length \(L\) will result in a systematic error in vertical component of the anchor position, but will be consistent for all vertical component estimates, hence its influence on displacement estimates is trivial (Appendix Figure A6). Note that thermal expansion may induce additional error in buoy length, but is small compared to the overall error budget and could be largely eliminated using a temperature-dependent model. At the test area, daily mean water temperature change between summer and winter is less than 25 °C (Appendix Figure A7). The thermal expansion coefficient for steel is about \(1.0 \times 10^{-5}\) m mm\(^{-1}\)K\(^{-1}\) (Okaji et al., 2000), thus thermal expansion-induced change in buoy length is on the order of a few mm.

Assuming reasonable errors for initial anchor coordinates in reference frame \(G-X_bY_bZ_b\) as 3/3/30 cm (i.e., \(p = \pm 3\) cm, \(q = \pm 3\) cm, \(\Delta L = \pm 30\) cm), and using the
repeatabilities provided by the digital compass manufacturer as errors for each heading/pitch/roll measurement (i.e., $\Delta \alpha = \pm 0.3^\circ$, $\Delta \beta = \pm 0.2^\circ$, $\Delta \gamma = \pm 0.2^\circ$ (Honeywell International Inc., 2019)), the maximum possible errors caused by errors in $p$, $q$, $L$, $\alpha$, $\beta$, and $\gamma$ are shown in Appendix formulas A8–A10. For a single epoch, errors in horizontal displacement estimates are at the centimeter level, and well below 1 cm for the vertical component. The standard deviations calculated for selected periods in Figure 2.12b are in approximate agreement with this theoretical analysis and hence are representative of the noise level in the daily displacement time series.

### 2.5.2 Response to environmental forcing

The GPS-buoy system rotates around the fixed anchor under the forcing of periodic tidal currents. Except for rapid motion during the first month after deployment due to settling, and the exceptionally rainy summer 2019, the anchor is relatively stable. Perturbations to anchor position occur mainly during extreme weather or periods of exceptionally high tidal current (Figures 2.11 and 2.13). These are analyzed below.

For the GPS and digital compass, the most prominent signal is the quasi-diurnal motion associated with ocean tides (Figures 2.6–2.8). Large pitch/roll values occur during high-speed current periods (Figure 2.8b and Supplementary Movies 1–3 in Xie et al. (2019c)). Figure 2.9 shows that most of the time the GPS phase center is located northwest of the estimated anchor position, indicating that outgoing tidal currents are stronger than incoming tidal currents. This reflects the geometry of Tampa Bay (Figure 2.1b), where fresh water from the Hillsborough River watershed adds to outgoing currents associated with the ocean tide. The geometry of the deep Egmont shipping channel focuses a direct outward tidal flow (jet) directly towards the buoy and the northern end of Egmont Key as it exits Tampa Bay (Figure 2.1b).

The buoy anchor moved westward by several meters immediately after deployment. The largest displacement occurred during the first two days after deployment. After that, large displacements correlate with periods of extreme weather conditions (Figures 2.11, 2.13
Figure 2.13: Anchor displacement and environmental variables during Hurricane Michael (a) and an unnamed storm in December 2018 (b). Water levels were observed at the site shown by the orange square in Figure 2.1b. Currents (Chen et al., 2018, 2019) are the Tampa Bay Coastal Ocean Model (TBCOM) hindcast at the GPS-buoy site. Observed currents were recorded by a current meter shown by the orange pentagon in Figure 2.1b. Wind speed and air pressure data were observed at the meteorological station shown by orange triangle in Figure 2.1b. Note that light blue shade marks the time of large displacement when current speed is at local maximum. Appendix Figure A8 shows anchor displacement and current along anchor long-term motion direction and local mean current direction, both showing that large anchor displacements occurred when current speeds are at local maximum.

and Appendix Figures A4, A8). The USGS coastal database shows that seafloor at the anchor location to be primarily sand-based (usSEABED database: https://coastalmap.marine.usgs.gov/js_map/national/usseabed/, also see Figure 2.2b). A bottom cavity (Figure 2.3) at the bottom of the ballast was designed to minimize translation movements on such
bottom types, which should give a friction coefficient \((f)\) of \(\sim 0.7\) (IALA Guideline 1066, 2010, and http://www.geotechdata.info/, last access on 4 February 2020). Using a software mechanical model (Working ModelTM), and assuming a wind speed of 20–25 m/s plus an ocean flow speed of 1.1–1.8 m/s during extreme weather conditions (e.g., the unnamed storm around December 20, 2018), motion of the ballast is unlikely if the ballast sits on a gentle slope (ballast tilt angle less than several degrees), due to a residual frictional force of \(>1.0 \times 10^4\) N (Appendix Figure A9). Models with much lower friction \((f = 0.3)\) suggest that ballast motion is possible under conditions of high tidal current forcing (Appendix Figure A9b). Additional modeling results suggest that the optimum site condition for the present design is a dense sand bed with slope smaller than 5° and current speed less than 2 m/s.

Current-induced scour changes conditions at the buoy anchor and is the most likely cause of the displacements at the test site. The left column of Figure 2.14 shows hillshade maps of the GPS-buoy site derived from multibeam sonar surveys. Large troughs formed near the anchor, probably contributing to anchor instabilities. The multibeam dataset used to choose the buoy location from April 2018 was collected with a Reson SeaBat T50-R dual-head 200kHz-400kHz system, run at 400kHz. The September 2018 dataset was collected with a Reson SeaBat 7125, also run at 400 kHz. All following multibeam datasets were collected with the Reson SeaBat T50-R dual-head run at 400 kHz. The bathymetry data derived from multibeam surveys have grid sizes of 0.25–0.5 m, depending on the density of sounding measurements. All maps are georeferenced in the same local reference frame. Geolocations of these bathymetry maps (with MarineStar) have an accuracy of about 0.1 m (https://www.fugro.com/our-services/marine-asset-integrity/satellite-positioning/marinestar). Comparison of multibeam sonar soundings of the anchor in the 27 September 2018 surface to the 20 March 2019 surface shows a southwestern displacement of 1–2 m, consistent with the anchor position estimates using the GPS-buoy system (Figure 2.14).
Figure 2.14: Bathymetry from multibeam sonar surveys showing bottom changes after buoy deployment. Hillshade maps (left column) are illuminated by a light source from west (azimuth angle is 270°, clockwise from north), with an elevation angle of 30°. White annotations on the lower right of (a, b, c, d) show dates of surveys. Details of the outlined cyan box are shown on the right column. In (a1, b1, c1, d1), background color maps show relatively heights with color bars changing from dark purple (deeper) to yellow (shallower), colored dots show daily median estimate of anchor positions with color bar change from wheat-green-black (located at the bottom of (d1)). In (b1, c1, d1), grey dots show 15-second interval GPS positions on corresponding days, centers of red “+” mark anchor position estimates for corresponding days. Note that bathymetry artifact in (c) is due to waves caused by ship turning, white pixels in (b1, c1, d1) are data gaps due to low density of valid sounding measurements.
Diver observations on October 1, 2019 support this view. They show the anchor sitting in a 1 meter deep scour hole, with a gap at one corner between the base of the anchor and the sediment-water interface. This suggests that the anchor periodically shifts after periods of high current speeds and scour activity, settling into a new position as scour activity changes the local slope beneath the anchor.

Most applications of the buoy would not involve locations in a tidal channel, so these kinds of perturbations would not likely occur. For applications where high tidal current speeds are expected, modification of the ballast design may be required.
3. Continued Development and Potential Applications

3.1 Continued development

3.1.1 Motivation

The system described in chapter 2 uses a semi-rigid connection from the GPS to the sea floor anchor. The design is suitable for water shallower than \( \sim 40 \) m. Its predecessor design, the MEDUSA system built by INGV (De Martino et al., 2014; Iannaccone et al., 2018), is capable of working in water as deep as 150 m. Based on the instrumentation and data analysis experience, we feel an extension of our current design to enable the system working in depth greater than 40 m is possible. A mooring cable will be used to connect the spar and the anchor, similar to the MEDUSA system. Considering the technical difficult, we aim at developing an updated version of the GPS-buoy system that can work in water up to 150 m deep. Figure 3.1 shows that in Cascadia and Central America, water depth shallower than this covers approximately 25% – 50% of the submerged areas. Compared to the existing system, an extension of the design that is able to work in water up to 150 m deep will greatly increase the capable spatial coverage.

Except for broader coverage, reduced cost is another benefit for using a mooring cable to connect the spar and the ballast. With the current design, the spar used to transmit GPS measurements to seafloor position must be longer than water depth, making engineering and transportation difficult and costly. In addition, using a mooring cable may make it possible to standardize the instrumentation for future network deployment. Currently, sizes of the spar, float, and anchor must be well coordinated to maintain the system in a stable and nearly vertical position. This means that customized design is required for each system. An
upgraded design with a mooring cable will reduce engineering effort since the cable length is the only main component that needs to be adapted for different depths.

Figure 3.1: Bathymetry of subduction zones in Cascadia (a) and Central America (b). Dashed magenta shapes in (a) and (b) outline areas used to calculate percentages of water coverage. Cyan and black arrows mark coast line distances of locations with 40 m and 150 m water depth, respectively. (c) shows percentages of area coverage for water shallower than the maximum depth. In Cascadia, approximately 25% of the submerged area between the coast line and trench is less than 150 meters in depth. In the Nicaraguan section of Central America, the corresponding figure is 50%. Note arrows in Figure 3.1 point to potential locations for future deployment since they are generally shallower than surrounding areas.
Figure 3.2: Sketch of proposed system with a mooring line, not to scale. Size and relative positions of the ballast, mooring cable, spar and float are adapted according to water depth and anticipated sea state. (a) shows that the system is deflected by ocean current from a vertical position, curvature of the cable is minimum compared to its length. In (b), curvature of the mooring cable is exaggerated for displaying propose.

3.1.2 Proposed design of a system capable for deeper water

Figure 3.2 illustrates the concept of design that works for water depth up to 150 m. A mooring cable is used to connect the spar buoy with anchor. The above waterline components will remain the same as in the current version. A taut wire (yellow line in Figure 3.2) will be used as mooring cable. Length of the cable is selected in accordance with water depth, so that the float will remain several meters below the water surface, and the superstructure will remain several meters above the water surface. The overall principle is
to ensure the entire system in a semi-rigid condition, so that position of the anchor can be resolved with GPS measurements.

Using a mooring cable in the proposed design will increase the degree of freedom for the system, making the horizontal position recovering more challenging. To assess the influence of adding a cable, I model the change of effective cable length in different scenarios, shown in Figure 3.3. An increase in net buoyancy provided by the float can reduce the influence of cable deflection due to ocean current. However, a larger float is more costly and requires a larger ballast weight. Thus it is important to evaluate size of the float based on the anticipated sea state and logistic ability.

![Figure 3.3: Change in effective length of a mooring cable. In (a), change of effective length is due to ocean current deflecting the mooring cable from a straight line (illustrated in Figure 3.2b). In (b), change of effective length is due: 1) ocean current deflecting the buoy from a vertical position, and 2) elongation of the cable due to elastic stretching caused by the buoyancy applied on the system. These calculations are for a standard 100 meter wire rope, and assume the current direction has no vertical component, current direction and cable path remain coplanar, current speed does not vary with depth, and the cable is in static equilibrium (not accelerating or decelerating). Mooring line models are based on Berteaux (1976). I assume the modulus of elasticity for the mooring cable is 200 GPa.

For places with high ocean current speeds, additional constraints may be needed to enable the anchor position be estimated with effective accuracy. For example, cable tilt and orientation may vary significantly along the the mooring line in highly variable sea
Sensors that can reliably measure cable tilt and orientation can potentially help better determining the anchor position. Such sensors may be integrated with the mooring cable.

### 3.2 Potential applications in shallow water seafloor geodesy

The experiment in Tampa Bay, Florida demonstrates that our GPS-buoy system is capable to detect centimeter seafloor displacement in shallow water and can operate autonomously. It has a number of applications in the shallow continental shelf environment and other active margins, where strain accumulation or release processes are poorly monitored. Below I list several examples of potential applications.

#### 3.2.1 Monitoring natural and human-induced offshore subsidence

In regions of offshore petroleum extraction, a sparse network could help define the subsidence pattern around oil and natural gas reservoirs as these products are extracted (Figure 3.4). Depleted offshore reservoirs have been proposed as possible sites for carbon capture and storage. Monitoring the surface deformation that is associated with these reservoirs as CO$_2$ is added can help to assess reservoir integrity (e.g., Karegar et al., 2015; Yang et al., 2015; Vasco et al., 2020).

The Mississippi Delta (Figure 3.4) has long been recognized as an area experiencing subsidence and land loss, due to a combination of natural and human-induced causes (e.g., Morton et al., 2006; Yuill et al., 2009). Compaction of younger Holocene sediments may be a significant cause (e.g., Meckel, 2008; Törnqvist et al., 2008). If dominant, then subsidence rate should correlate with the thickness of the underlying Holocene section (e.g., Figure 2a in Karegar et al., 2015). Some offshore areas in the delta can have considerably thicker Holocene sections compared to on-land areas, hence measuring offshore subsidence could help to better define the processes contributing to subsidence, assuming initial subsidence of the anchor during the settling period was accounted for.
3.2.2 Monitoring submerged volcanoes

Monitoring seafloor vertical displacement associated with volcano deformation using the MEDUSA system developed by INGV is well described in De Martino et al. (2014) and Iannaccone et al. (2018). In some volcano-tectonic regions, shallow water environments preclude precise measurement of strain or displacement fields. Lago Nicaragua in southern Nicaragua is an example of fresh water lake hosting volcanoes. The lake occupies more than 8,000 km², and limits measurement of surface deformation to the flanks of the volcanoes and the distant shorelines several kilometers from the volcanic islands. The majority of the lake is shallower than 40 meters depth, and hence is amenable to geodetic measurements with our current system. Marine examples include Anak Krakatoa in Indonesia (Figure 3.5), which caused a destructive tsunami that killed at least 437 people on 22 December 2018 (Ye et al., 2020). Because no large earthquake near the tsunami source was reported before the event,
there was no tsunami warning. Post-tsunami studies show that the tsunami was most likely caused by a subaerial landslide (flank collapse) during eruption of Anak Krakatoa (e.g., Muhari et al., 2019; Ye et al., 2020). Similar to Anak Krakatoa, many marine volcanoes are capable of producing tsunamis but not well monitored. The shallow water seafloor geodetic system can potentially be used to provide necessary geodetic coverage for them, especially at portions where submarine mass wasting events are likely to occur.

### 3.2.3 Monitoring subduction offshore strain processes

In subduction zones, some of the largest earthquakes occur offshore (e.g., 2011 $M_w \geq 9.0$ Tohokuoki earthquake). Offshore strain accumulation and release processes are critical for understanding megathrust earthquakes and tsunamis, but are usually poorly monitored. Parts of the Central America subduction system have a large shallow water section. In Nicaragua for example, nearly 50% of the area between the coastline and the trench is shallower than 150 meters (Figure 3.6a). In the Nicoya Peninsula of Costa Rica, Dixon et al. (2014) found that shallow slow slip events during the inter-seismic period release a significant amount of accumulated strain, perhaps reducing earthquake magnitude and tsunami potential. Similar results have been observed offshore Ecuador (Rolandone et al., 2018). Our GPS-buoy system is a promising geodetic tool to monitor seafloor motion in these shallow fore-arc regions.

Figure 3.6 is an example of how such a system could improve slip resolution of offshore events. Figure 3.6a shows the bathymetry of the fore-arc region in Central America. 10 cm dip slips were simulated on the plate interface (red patches). Figures 3.6 c–f show model resolution length scales with different datasets, based on Funning et al. (2005). Smoothing parameters used in Figures 3.6 c–f are the same, estimated using existing onshore cGPS data only. An elastic half-space model (Okada, 1992) is used to calculate Green functions, and the Tikhonov regularization method is used to maximize smoothing while minimizing residuals, similar to Malservisi et al. (2015). Even with numerous continuous GPS (cGPS) stations on land (Figures 3.6 c and d), slip distribution near the trench is not well resolved.
Figure 3.5: Bathymetry and radar images of the Anak Krakatau volcano. (a) Bathymetry near the 22 December 2018 tsunami source area. Red triangle marks the Anak Krakatau volcano. Black box outline the area shown in (b) and (c). Bathymetry data were downloaded from http://tides.big.go.id/DEMNAS/ (last access on 6 February 2020). (b) and (c) show satellite radar images before and after the tsunami. Note collapse of the flank and tsunami impacts in the red ellipse.

Adding a relatively small number of GPS-buoy stations at selected shallow locations would significantly improve the resolution of estimated slip distribution offshore (Figures 3.6 e and f).
Figure 3.6: Potential application area in Central America. (a) Bathymetry of the fore-arc region in Central America. Beach balls show $M_w > 7.5$ earthquakes between 1 January 1980 and 1 March 2019. Dashed line box outlines the 1992 Nicaragua earthquake tsunami fault model by Satake (1994). Solid line box outlines the seismological fault model by Ide et al. (1993), adopted from Figure 1 in Satake (1994). (b–f) Resolution test of slip inversion with/without seafloor geodetic measurements. Horizontal size of each small patch is 5 km × 5 km. Black dots are existing continuous GPS (cGPS) on land, green dots are proposed cGPS on land, black triangles are synthetic GPS-buoy sites. Purple lines mark location of the trench. Grey lines are slab depth contour (in km, Hayes et al. (2018)). Land GPS and offshore GPS sites are weighted with different uncertainties: land GPS 0.2/0.2/0.4 cm on north/east/vertical directions; offshore GPS 2.0/2.0/0.5 cm on north/east/vertical directions. (c) shows resolution length scale with existing land cGPS stations. (d) shows resolution length scale with existing land cGPS stations plus 8 synthetic land cGPS stations. (e) shows resolution length scale with land cGPS plus 4 synthetic offshore GPS-buoy stations (shallower than 40 m). (f) shows resolution length scale with land cGPS plus 8 synthetic offshore GPS-buoy stations (shallower than 100 m). For comparison, resolution remains unchanged in Nicoya Peninsula, Costa Rica (lower right), consistent with the model resolution of Kyriakopoulos and Newman (2016).
4. Conclusion

We have developed a GPS-buoy system suitable for measuring horizontal and vertical components of shallow water seafloor motion. The system consists of a GPS station, a digital compass, a spar buoy anchored to seafloor by a heavy ballast, and a float integrated into the spar. GPS data are processed in the kinematic mode, and a 3-dimensional transformation is used to estimate anchor position based on GPS position and buoy heading/pitch/roll measurements. During a test period in Tampa Bay, the system successfully recorded transient events associated with Hurricane Michael and other heavy precipitation events, when current speeds are high. For a single measurement, uncertainty induced by errors in buoy geometry and digital compass measurements is at the centimeter level. Using a median filter, the daily position time series of the anchor have a repeatability of 1–2 cm or better for the three position components. This system can be applied to a variety of study areas, including offshore regions of subduction zones for measuring strain accumulation and release processes. A combination of on-land GPS, GPS-buoys in shallow water, and GPS-acoustic systems or pressure gauges in deep water, would provide complete geodetic coverage for subduction zone earthquake, and tsunami and volcano hazard studies.
References


Appendices
Appendix I: Licence and reprint of Xie et al., 2016, JGlaciol

Copyright notice

The Journal of Glaciology is published as Gold Open Access with CC-BY licensing attribution 4.0 as default. There are other licencing attributions available.

If CC-BY is selected you are free to

Share — copy and redistribute the material in any medium or format

Adapt — remix, transform, and build upon the material

For any purpose, even commercially

Under the following terms

Attribution — You must give appropriate credit, provide a link to the license, and indicate if changes were made. You may do so in any reasonable manner, but not in any way that suggests the licensor endorses you or your use.

No additional restrictions — You may not apply legal terms or technological measures that legally restrict others from doing anything the license permits.

Please note that permissions are still required for using material published in IGS publications prior to 2016 when the publications became Gold Open Access.

To request a permission letter please e-mail the IGS at igsoc@igsoc.org

Note: as the copyright is jointly vested with both the author and the International Glaciological Society, you should also seek permission from the respective author of the paper you quote.
Advances in winter when the dense and strong ice mélange affect the seasonal variations in at Jakobshavn Isbrae to demonstrate that sea ice coverage and lapse imagery, GPS, ocean pressure and seismic observations calving models. Amundson and others (2010) used time-matics near the calving front can provide constraints on conditions. Detailed observations of ice geometry and kine-must be considered in calving criteria at least under certain major calving event, which implies that basal crevassing the front of Helheim Glacier tipped backwards during a obvious conflicts. A finite-element model of stress evolution near the front of marine-terminating glaciers sug-gests that undercutting of the ice front due to frontal melting near the base is a strong driver of calving (O’Leary and Christoffersen, 2013). However, a vertical 2-D ice flow model found that crevasses water depth and basal water pres-sure could have significant effects, while submarine melt undercutting and backstress from ice mélange are less im-portant (Cook and others, 2014). The models of Cook and others (2014) and many others (e.g., Nick and others (model CDw), 2010; Otero and others, 2010), use the calving criterion of Benn and others (2007), which assumes that calving happens when the depth of surface crevasses reaches the waterline, and does not require a basal crevas-sing condition. Recent work by Murray and others (2015a, b) cast doubt on this calving criterion. Their data show that the front of Helheim Glacier tipped backwards during a major calving event, which implies that basal crevasseing must be considered in calving criteria at least under certain conditions. Detailed observations of ice geometry and kine-matics near the calving front can provide constraints on calving models. Amundsson and others (2010) used time-lapse imagery, GPS, ocean pressure and seismic observations at Jakobshavn Isbrae to demonstrate that sea ice coverage and the strength of mélange affect the seasonal variations in calving rate and terminus stability: the glacier terminus advances in winter when the dense and strong ice mélange prevents calving, and retreats in summer when the ice mélange becomes weak. A simple force-balance analysis suggested that where there is a resistive ice mélange, bottom-out rotation of the calving block is strongly preferred over top-out rotation. By using photogrammetric time-lapse imagery, Rosenau and others (2013) documented a major calving event at Jakobshavn Isbrae, finding large vertical displacements of the glacier front of order 15 m and lowering of order 8 m at a position 500 m from the calving front 2 d before the calving event, similar to the observations at Helheim Glacier by Murray and others (2015a, b).

Terrestrial radar interferometry (TRI) allows detailed obser-vations of the calving front, generating high-resolution elevation and velocity data with short (several minutes or less) repeat intervals (Dixon and others, 2012; Peters and others, 2015; Voytenko and others, 2015a, b, c). With this instrument, we can measure glacier motion and map ice velocity and elevation over a wide area, overcoming the limitations of GPS (low spatial resolution, difficult to deploy near the calving front), photogrammetry (low reliability in bad weather and at night), and satellite observations (low temporal resolution). Using continuous TRI observations near the terminus of Jakobshavn Isbrae acquired for 4 d in June 2015, we discuss the possible role of crevasses and basal melting before and during a calving event.

1. INTRODUCTION
Jakobshavn Isbrae, Greenland’s largest marine-terminating glacier, has doubled in speed as its ice front has retreated tens of km in the last several decades (Morgan and others, 2004, 2008; Rignot and Kanagaratnam, 2006; Howat and others, 2011). Increases in subsurface melting and calving triggered by warmer ocean water are believed to be important contributors to this process (Holland and others, 2008; Motyka and others, 2011; Enderlin and Howat, 2013; Myers and Risergaard, 2013; Truffer and Motyka, 2016).

Modeling the calving process is challenging, and has produced conflicting results. A finite-element model of stress evolution near the front of marine-terminating glaciers suggests that undercutting of the ice front due to frontal melting near the base is a strong driver of calving (O’Leary and Christoffersen, 2013). However, a vertical 2-D ice flow model found that crevasse water depth and basal water pressure could have significant effects, while submarine melt undercutting and backstress from ice mélange are less important (Cook and others, 2014). The models of Cook and others (2014) and many others (e.g., Nick and others (model CDw), 2010; Otero and others, 2010), use the calving criterion of Benn and others (2007), which assumes that calving happens when the depth of surface crevasses reaches the waterline, and does not require a basal crevasseing condition. Recent work by Murray and others (2015a, b) cast doubt on this calving criterion. Their data show that the front of Helheim Glacier tipped backwards during a major calving event, which implies that basal crevasseing must be considered in calving criteria at least under certain conditions. Detailed observations of ice geometry and kinematics near the calving front can provide constraints on calving models. Amundsson and others (2010) used time-lapse imagery, GPS, ocean pressure and seismic observations at Jakobshavn Isbrae to demonstrate that sea ice coverage and the strength of mélange affect the seasonal variations in calving rate and terminus stability: the glacier terminus advances in winter when the dense and strong ice mélange prevents calving, and retreats in summer when the ice mélange becomes weak. A simple force-balance analysis suggested that where there is a resistive ice mélange, bottom-out rotation of the calving block is strongly preferred over top-out rotation. By using photogrammetric time-lapse imagery, Rosenau and others (2013) documented a major calving event at Jakobshavn Isbrae, finding large vertical displacements of the glacier front of order 15 m and lowering of order 8 m at a position 500 m from the calving front 2 d before the calving event, similar to the observations at Helheim Glacier by Murray and others (2015a, b).

Terrestrial radar interferometry (TRI) allows detailed obser-vations of the calving front, generating high-resolution elevation and velocity data with short (several minutes or less) repeat intervals (Dixon and others, 2012; Peters and others, 2015; Voytenko and others, 2015a, b, c). With this instrument, we can measure glacier motion and map ice velocity and elevation over a wide area, overcoming the limitations of GPS (low spatial resolution, difficult to deploy near the calving front), photogrammetry (low reliability in bad weather and at night), and satellite observations (low temporal resolution). Using continuous TRI observations near the terminus of Jakobshavn Isbrae acquired for 4 d in June 2015, we discuss the possible role of crevasses and basal melting before and during a calving event.

2. DATA ACQUISITION
We observed the terminus of Jakobshavn Isbrae with a TRI from June 6–10 2015. The instrument is a real-aperture radar operating at Ku-band (1.74 cm wavelength) and is sensitive to line-of-sight (LOS) displacements of ~1 mm (Werner and others, 2008). The instrument was mounted on a metal pedestal on solid rock ~3 km away from the calving front, and protected by a radome to eliminate disturbance from wind and rain (Fig. 1). Figure 2 shows the area measured during 4 d of continuous observation. The TRI scanned a 150° arc at a sampling rate of 90 s, generating images with both phase and
The resolution of the range measurements is ∼1 m. The azimuth resolution varies linearly with distance: for example, 7 m at 2 km distance, 14 m at 4 km.

The TRI has one transmitting antenna and two receiving antennas, which allow for repeat topographic mapping of fast moving glaciers (Strozzi and others, 2012; Voytenko and others, 2015a). The baseline length (vertical offset between the two receiving antennas) in this campaign was 60 cm.

3. DATA ANALYSIS AND RESULTS

We first converted unwrapped phases into elevation maps using a geodetic reference height on the stationary rock, then adjusted the elevation into a local height coordinate system relative to the mean water level in the fjord. The results were resampled into 10 m pixel spacing maps and georeferenced into UTM coordinates for further analysis.

The TRI captured several small calving events during its 4-d observation period, and one large calving event near the end. Here we focus on the large calving event. Figure 3 shows the intensity images before (a) and after (b) this event. Surface dimensions of the calved block are ∼1370 m × 290 m.

For fast moving glaciers like Jakobshavn Isbræ, ice near the terminus can move over 30 m d⁻¹, so the location of the calving front can change more than 120 m during 4 d of observation. This motion must be considered when analyzing elevation variations of the glacier front. Our radar data are acquired in a fixed Cartesian system, so a given ice particle at the surface of the glacier travels through this Cartesian coordinate system (Eulerian reference frame). For this study, it is also useful to consider a Lagrangian reference frame, where we track a given particle of ice through time.

We converted our elevation time series, originally defined in an Eulerian frame, into a Lagrangian frame, as follows:

\[
H_{\text{Lag}}(x_{\text{Lag}}, y_{\text{Lag}}) = H_{\text{Eul}}(x_0 + dx, y_0 + dy) \quad (1)
\]

where \(H_{\text{Lag}}\) and \(H_{\text{Eul}}\) are elevations in the Lagrangian and Eulerian frame, respectively; \((x_{\text{Lag}}, y_{\text{Lag}})\) are the coordinates in the Lagrangian system, set equal to the initial coordinates \((x_0, y_0)\) at \(t_0\) in the Eulerian frame; and \(dx\) and \(dy\) are the horizontal displacements (relative to \(t_0\)) of ice at time \(t\) in the Eulerian frame.

To obtain \(dx\) and \(dy\) in Eqn (1), we estimated ice motion by using the feature tracking method in OpenCV (http://opencv.org/). Figure 4 is an example of ice motion derived by
tracking distinct features such as the edges of surface crevasses on TRI intensity images. The velocity of the ice mélange is quite variable since even small calving events can cause large mélange motion. The glacier motion is variable over hourly timescales, but is relatively consistent over longer (1 d) periods. The estimated speed near the calving front was $\sim 34 \text{ m d}^{-1}$ during our observations.

Topographic mapping with the TRI is based on the interferometric imaging geometries of the two receiving antennas and the various targets in the imaged swath (Strozzi and others, 2012). Two steps are necessary to convert unwrapped phases into elevation maps. First, we need to estimate the ‘expected’ phase at the radar position based on the elevation difference between the instrument and the reference point. Second, an elevation map is derived from the phase difference between the ‘expected’ radar phase and the unwrapped phase map. Ideally, for the first step, if we choose a stationary point (e.g., rock) as the reference elevation point, the ‘expected’ phases of the radar at different times should be the same. In reality, however, the phase of the radar position estimated at different times can be slightly different because of measurement noise. Since we hope to exploit the time-varying DEM capability of our TRI instrument, we cannot rely on long time (hour-scale) averages of multiple DEMs to reduce random noise in the elevation estimates. This noise is mainly due to atmospheric propagation effects (especially from variable water vapour) and possible small variations in antenna orientation associated with the scanning motion of the radar (the radome eliminates antenna motion due to wind).

We corrected the elevation estimates in two stages. As described above, a first order correction is applied by subtracting the ‘expected’ phase differences from a stationary point on rock $\sim 600$ m away from the instrument. This corrects the majority of effects due to antenna wobble, but may not improve the elevation estimates in the areas of interest on the glacier, as these are farther from the radar, and the radar signal propagates through atmosphere that is spatially and temporally variable. In a second step, we use elevation estimates in the mélange immediately in front of the glacier (box b in Fig. 5a) to correct the DEM on the glacier near the calving front, since we expect noise sources in the two areas to be similar. Tidal signals in the mélange are of order 1 m in amplitude, below the noise level of the elevation estimates, and we assume that over the 4 d of observation, the mean elevation change in this area is close to zero (no large icebergs entered the area during this period until the studied calving event). The resulting RMS scatter in the mélange (box b) is 1.8 m (Fig. 5b), about the level expected given instrument noise, atmospheric effects and tides. The elevations on the nearby glacier change by amounts that are much larger, but have several ‘tears’ in the time series associated with phase breaks. The deviations from the mean height in the mélange are used to correct these phase

![Fig. 4. Daily ice velocity estimated by tracking motion of distinct features. Blue boxes in (a) outline areas shown in more detail in (b), (c). Length of arrows is on the same scale as the background TRI intensity images (they are in the same reference coordinate system, 1 pixel length = 10 m). Black areas are in radar shadow.](image-url)
The corrected elevations still exhibit changes across the glacier front that are up to an order of magnitude higher than the changes in mélange (Fig. 5c).

Figure 5a is an averaged elevation map overlain on a Landsat-8 image. Figure 5c shows the corrected elevation profiles of points separated by 10 m along an approximate flow line beginning near the cliff that calved during the main calving event. The black arrow indicates the time of calving on 10 June 2015. In the 4 d before the large calving event the elevation of the glacier front increased by up to 20 m, in a way that is consistent with a simple block rotation model, as described below. These results are similar to the findings of Murray and others (2015a) who studied Helheim Glacier with GPS and photogrammetry. They suggested that glaciers can calve by a process of buoyancy-induced crevassing, with ice down-glacier in zones of flexure rotating upward (bottom-out rotation) because of disequilibrium.

The pattern of elevation increases along a flow line close to the ice cliff can be explained as follows. Assuming block behaviour, as the frontal ice block begins to flex at the beginning of a calving event, elevations near the cliff initially increase and the basal crevasse evolves and widens. Once the ice block is significantly out of equilibrium, ice failure can happen rapidly. The ice flexure and crevasse growth are separate physical processes that can be mutually
reinforcing. We simplified these processes with a model of a single rigid block undergoing rotation with no internal deformation. Figure 6 is a cartoon showing the process. The new TRI data allow us to describe the timing and geometry of this process in some detail.

The surface width of the ice block, \( W \), can be determined directly from the TRI intensity images before and after the calving event. On a cross section plane, for a point on the upper surface with initial distance of \( d_0 \) to the ice cliff, and initial elevation of \( h_0 \), the horizontal distance from this point to the cliff is:

\[
d = W - [(W - d_0) \cos \theta - (h_0 - H_0) \sin \theta]
\]

where \( H_0 \) is the initial height of the intersection axis (the top of the calving surface of Fig. 6a) and \( \theta \) is the rotation angle. The expression for elevation is:

\[
h = (W - d_0) \sin \theta + (h_0 - H_0) \cos \theta + H_0 - D
\]

where \( D \) is the downward motion of the ice block (Fig. 6). Equations (2) and (3) assume the ice block rotates about the intersection axis in a rigid way.

To test this model, we selected a profile along a line that is perpendicular to the calving surface (the angle between the profile and the flow line direction is \( \sim 33^\circ \)), and estimated elevations along the profile at different times (Fig. 7). Our time-varying DEMs effectively represent 15 min time averages, and are generated as follows: For each time increment, we derive elevations from five scans on each side (total of ten scans, spanning 15 min) and take the median value. If there are no usable measurements within a given 15 min increment, then there are no elevation estimates for that time. For comparison, different colour-coded curves in Figure 7 show the best-fit estimates for a rigidly rotating ice block, allowing the block edge on the upstream side to shift downward on the new ice cliff as calving proceeds. Figure 8 plots the rotation angle as a function of time (Supplementary Fig. S1 plots the downward motion as a function of time). Note the sudden drop at \( \sim 28.5 \) h before the calving event, which coincides with the time that a small piece of ice on the edge of the calving block fell from the cliff (Fig. 9).

Rosenau and others (2013) used time-lapse photography to suggest that vertical displacements of the glacier front at Jakobshavn Isbræ began \( \sim 2 \) d before a calving event. From Figure 7 and 8, we conclude that the calving process for our studied event actually started at least \( \sim 65 \) h prior to the visually observed calving event.

4. DISCUSSION

Our simple block rotation model describes glacier front motion several days prior to a major calving event. The model has just three parameters: block width (\( W \)), rotation angle (\( \theta \)) and downward motion of the up-glacier edge of the block (\( D \)). \( W \) is determined directly from the intensity images, while \( \theta \) and \( D \) are determined by fitting the elevation time series data with model predictions. Figures 7, 8 show that the ice block started rotating at least \( \sim 65 \) h before the calving event, a clear strain precursor to subsequent ice failure. The cross-over points in Figure 7 define an approximate lower bound for the width of the future calved block. Figure 7 also suggests that the elevation of ice close to the rotation axis decreases during the later stages of the calving process. The TRI intensity images support this: the observable ice surface becomes narrower as the up-glacier ice subsides and is shadowed by the higher down-glacier ice (Supplementary Fig. S2).
4.1. The role of subsurface melting

By studying tidal responses with photogrammetric time-lapse images, Rosenau and others (2013) found a narrow floating zone near the frontal cliff of Jakobshavn Isbræ. Our TRI-derived ice velocity estimates and phase lags relative to ocean tides suggest a ∼1 km wide floating zone near the terminus during the observation period (Supplementary Information). The studied calving event happened at the front of this zone. Ice in the floating zone experiences tidal flexing, which can initiate Mode 1 (opening) cracks. These can form as both surface and basal crevasses. While surface crevasses can grow rapidly during a summer, basal crevasses can probably grow more rapidly if warm water is circulating in the fjord, reflecting the higher heat capacity of water relative to air.

Luckman and others (2015) suggest that glacier undercutting driven by warm ocean temperature is an important process that contributes to calving in marine-terminating glaciers. We hypothesize that at the floating zone, where subsurface melting is likely faster than surface melting, the ice block moves out of gravitational equilibrium, which flexes the ice in a narrow zone (within which the new calving
As ice in this narrow zone flexes and the block rotates, crevasses enlarge, deforming zones narrow and strain increases exponentially. Eventually a failure threshold is reached and the block collapses. Figure 10 sketches the process. This model also explains the step change in elevation and rotation angle ∼28.5 h before the calving event (Fig. 8): the preliminary ice fall removed mass above the water line, allowing the block to temporarily rebound. Continued subsurface melting eventually allowed the process to continue.

We can test this hypothesis by considering the differential stress generated by plausible amounts of subsurface melting, and comparing with laboratory-measured strength of ice. This analysis (see Supplementary Information) suggests that losses of order 30% are required to generate buoyancy-related differential stresses sufficient to initiate failure. This seems high, although ice in the terminal zone may be significantly weaker than laboratory-derived values, depending on the depth of pre-event crevassing. Perhaps a combination of surface and basal crevassing is necessary.

4.2. Ice failure model

Voight (1988) described a method for predicting material failure in rocks, soil and other solids under stress:

\[ \dot{\Omega} = \frac{\alpha}{C_0} \]

where \( A \) and \( \alpha \) are empirical constants, \( \Omega \) is an observable quantity related to deformation, and one and two dots refer to the first and second derivatives with respect to time. We suggest this model can also be applied to calving ice. We applied the model to the rotation of the ice block at Jakobshavn Isbræ, with \( \Omega \) taken as the rotation angle \( \theta \), and the rotation rate \( \dot{\theta} \) assumed to be infinite at the time of calving. By using a grid search approach, \( A \) and \( \alpha \) were estimated to be 23.4 and 4.5. Following Voight (1988), the expression for rotation rate when \( \alpha > 1 \) is:

\[ \dot{\theta} = \frac{A(\alpha - 1)}{(t - t_f) + \theta^2(t - t_f)} \left(\frac{t}{1 - \alpha}\right) \]

where subscript \( f \) indicates the time of failure. Figure 11 shows the block rotation rate versus time. The weighted Fig. 10. Sequential sketches of the physical process for calving. (a) Ice near calving front is neutrally buoyant. (b) Submarine melting exceeds surface melting, hence the ice block is no longer gravitationally stable. (c) Ice block sinks and rotates, basal crevasse enlarges, and the block eventually calves.

Fig. 11. Ice block rotation rate (red dots) versus time. At time 0 the iceberg collapses and we assume the rotation rate is infinite. Grey curve is the best fit of ice failure model with \( A \) and \( \alpha \) equal to 23.4 and 4.5, respectively. WRMS residual of model fit is 0.07° h\(^{-1}\); weights of rates are based on misfits of the rotation model shown in Figure 7. Rotation rate estimates are based on rotation angles shown in Figure 8, using a least-squares smoothing filter (Gorry, 1990), with smoothing window =5 and local polynomial approximation of order =2. Note that the model fits both the rotation rate data as well as the calving time data.
locations and elevations are computed from Eqns (2) and (3). The rotation angles and downward motions (or downward motion) can be expressed as (Voight, 1988):

\[ \theta = \frac{1}{A(a - 2)} \left\{ \left[ A(a - 1)\left(t_1 - t_0\right) + \delta_1^{\left(a-2\right)}\right]t_1^{-\left(a-1\right)} - \left[ A(a - 1)\left(t_0 - t\right) + \delta^{\left(a-2\right)}\right]t^{-\left(a-1\right)} \right\} \]  

(6)

We add a Heaviside step function to account for the small ice failure event ∼28.5 h before the main calving event. For downward motion, the values of A and a are estimated to be 1.1 and 4.3. Based on these estimates and the model, we can derive the elevations of selected points at time t before the calving event. Figure 12 is a plot of TRI-derived elevations and predictions based on the ice failure and block rotation models. The rotation angles and downward motions are computed by Eqn (6), assuming \( t_0 = -80 \) h. The profile locations and elevations are computed from Eqqs (2) and (3). The ice failure model parameters are sensitive to observations immediately before the calving event. Due to limited data quality, the uncertainties of the model predictions are relatively high. Note that the model ignores tidal forcing and assumes no internal deformation in the calving block. Analysis of tidal variations in the fjord shows no evidence that block rotation or ice failure are sensitive to tide or tide rate, but tidal flexing of the ice in the floating zone could extend the depth of crevasses and fracture and weaken the ice (see Supplementary Information). Improved precision in the time-varying elevation estimates, better estimates of local tides, and detailed block shape variations should allow for a better understanding of ice flexure during calving.

5. CONCLUSIONS

We used TRI-derived digital elevation models to investigate the behaviour of the calving front at Jakobshavn Isbrae. Ice elevation near the cliff began to increase several days before a major calving event on 10 June 2015. A simple rigid block rotation model matches the elevation profiles and suggests that block capsizing started ∼65 h prior to calving. Subsurface melting in excess of surface melting may over-weight the above-water mass of ice and enhance crevasses, leading to ice deformation, block rotation and eventual ice failure. A simple failure model fits the rotation data quite well.

SUPPLEMENTARY MATERIAL

The supplementary material for this article can be found at http://dx.doi.org/10.1017/jog.2016.104.

ACKNOWLEDGEMENTS

This research was partially supported by NASA grant NNX12AK29G to THD. DMH acknowledges support from NYU Abu Dhabi grant G1204 and NSF grant ARC-130413.7. We thank Martin Truffer and an anonymous reviewer for their valuable comments.

REFERENCES


Cook S and 7 others (2014) Modelling environmental influences on calving at Helheim Glacier in eastern Greenland. Cryosphere, 8, 827–841 (doi: 10.5194/tc-8-827-2014)


Xie and others: Precursor motion to iceberg calving at Jakobshavn Isbræ observed with terrestrial radar interferometry

MS received 1 April 2016 and accepted in revised form 2 August 2016; first published online 19 September 2016

65
Appendix II: Licence and reprint of Xie et al., 2018, TC

Licence and copyright agreement

Author’s certification

By submitting a manuscript, the authors certify that they have read and agreed to the following terms:

- The authors are authorized by their co-authors to enter into these arrangements.
- The work is original and has not been formally published before (except in the form of an abstract, preprint, or as part of a published lecture, review, or thesis), that it is not under consideration for publication elsewhere, that its publication has been approved by all the author(s) and by the responsible authorities – tacitly or explicitly – of the institutes where the work has been carried out, and that the article does not infringe copyright or any other rights by third parties.
- The work does not contain content that is unlawful, abusive, or constitute a breach of contract or of confidence or of commitment given to secrecy.
- The authors warrant that they secure the right to reproduce any material that has already been published or copyrighted elsewhere and that they identified such objects with appropriate citations and copyright statements, if applicable, in captions or even within the objects themselves (e.g. copyrights of maps).
- They agree to the following licence and copyright agreement:

Copyright

- Authors retain the copyright of the article. Regarding copyright transfers please see below.
- Authors grant Copernicus Publications an irrevocable non-exclusive licence to publish the article electronically and in print format and to identify itself as the original publisher.
- Authors grant Copernicus Publications commercial rights to produce hardcopy volumes of the journal for sale to libraries and individuals.
- Authors grant any third party the right to use the article freely as long as its original authors and citation details are identified.
- The article is distributed under the Creative Commons Attribution 4.0 License. Unless otherwise stated, associated published material is distributed under the same licence.

Creative Commons Attribution 4.0 License

You are free to:

- Share — copy and redistribute the material in any medium or format
- Adapt — remix, transform, and build upon the material for any purpose, even commercially.

Under the following conditions:

- Attribution — You must give appropriate credit, provide a link to the licence, and indicate if changes were made. You may do so in any reasonable manner, but not in any way that suggests the licensor endorses you or your use.
- No additional restrictions — You may not apply legal terms or technological measures that legally restrict others from doing anything the licence permits.

It is important to note that the Creative Commons Attribution 4.0 License and the Open Government License (OGL) are interoperable and do not conflict with, reduce or limit each other.

© Crown copyright YEAR

Appendix II: Licence and reprint of Xie et al., 2018, TC

TC - Licence & copyright

The works published in this journal are distributed under the Creative Commons Attribution 4.0 License. This licence does not affect the Crown copyright work, which is re-usable under the Open Government Licence (OGL). The Creative Commons Attribution 4.0 License and the OGL are interoperable and do not conflict with, reduce or limit each other.

© Crown copyright YEAR
Reproduction request

All articles published by Copernicus Publications have been licenced under the Creative Commons Attribution 4.0 License since 6 June 2017 or under its former version 3.0 since 10 December 2007. Under these licences the authors retain the copyright. There is no need from the publisher’s side to allow/confirm a reproduction. We suggest contacting the authors to inform them about the further usage of the material. In any case, the authors must be given credit. If articles contain figures, maps, or other objects cited by the authors, the individual copyrights and distribution licences must be clarified individually.
Grounding line migration through the calving season at Jakobshavn Isbræ, Greenland, observed with terrestrial radar interferometry

Surui Xie¹, Timothy H. Dixon¹, Denis Voytenko², Fanghui Deng¹, and David M. Holland²,³

¹School of Geosciences, University of South Florida, Tampa, FL, USA
²Courant Institute of Mathematical Sciences, New York University, New York, NY, USA
³Center for Global Sea Level Change, New York University, Abu Dhabi, UAE

Correspondence: Surui Xie (suruixie@mail.usf.edu)

Received: 10 October 2017 – Discussion started: 5 January 2018
Revised: 17 March 2018 – Accepted: 27 March 2018 – Published: 17 April 2018

Abstract. Ice velocity variations near the terminus of Jakobshavn Isbræ, Greenland, were observed with a terrestrial radar interferometer (TRI) during three summer campaigns in 2012, 2015, and 2016. We estimate a ~1 km wide floating zone near the calving front in early summer of 2015 and 2016, where ice moves in phase with ocean tides. Digital elevation models (DEMs) generated by the TRI show that the glacier front here was much thinner (within 1 km of the glacier front, average ice surface is ~100 and ~110 m above local sea level in 2015 and 2016, respectively) than ice upstream (average ice surface is >150 m above local sea level at 2–3 km to the glacier front in 2015 and 2016). However, in late summer 2012, there is no evidence of a floating ice tongue in the TRI observations. Average ice surface elevation near the glacier front was also higher, ~125 m above local sea level within 1 km of the glacier front. We hypothesize that during Jakobshavn Isbræ’s recent calving seasons the ice front advances ~3 km from winter to spring, forming a >1 km long floating ice tongue. During the subsequent calving season in mid- and late summer, the glacier retreats by losing its floating portion through a sequence of calving events. By late summer, the entire glacier is likely grounded. In addition to ice velocity variation driven by tides, we also observed a velocity variation in the mélange and floating ice front that is non-parallel to long-term ice flow motion. This cross-flow-line signal is in phase with the first time derivative of tidal height and is likely associated with tidal currents or bed topography.

1 Introduction

Greenland’s largest marine-terminating glacier, Jakobshavn Isbræ, has doubled in speed and retreated tens of kilometers in the last few decades (Joughin et al., 2004, 2008; Rignot and Kanagaratnam, 2006; Howat et al., 2011). This process has been attributed to several processes, including increased subsurface melting and iceberg calving triggered by relatively warm ocean water (Holland et al., 2008; Motyka et al., 2011; Enderlin and Howat, 2013; Myers and Ribergaard, 2013; Truffer and Motyka, 2016). In recent years, the glacier has maintained a relatively stable terminus position despite continued speedup, primarily due to the fact that the glacier is now embedded in the ice sheet, with large inflows of ice from the sides supplying ice to the main glacier channel, albeit with some thinning (Joughin et al., 2008). However, it is not clear if this configuration is stable, as Jakobshavn Isbræ has a retrograde bed (Clarke and Echelmeyer, 1996; Gogineni et al., 2014). Some numerical models suggest that glaciers with reverse bed slopes cannot maintain stable grounding lines, as bed topography favors ingress of warm fjord bottom water, accelerating melting at the ice–ocean interface (e.g., Vieli et al., 2001; Schoof, 2007).

In addition to the dramatic secular speedup and retreat, there are strong seasonal variations in both ice speed and front position at Jakobshavn Isbræ. These have an inverse correlation: ice accelerates through spring and summer but slows down in winter, while glacier front position retreats from spring to summer, reaching a minimum in late summer when ice speed is maximum (Joughin et al., 2008). This supports the hypothesis that loss of the buttressing ice tongue during the calving season contributes to Jakobshavn
Isbær’s seasonal speedup. The rapid acceleration since 2000 may thus be the sequential result of losing its large floating ice tongue from 1998 to 2003 (Joughin et al., 2004, 2008), though Van Der Veen et al. (2011) suggested that progressive weakening of ice in the lateral shear margins is a more plausible explanation for the acceleration. By investigating interactions between the glacier and its pro-glacial ice mélange, Amundson et al. (2010) interpreted the seasonal advance and retreat of the glacier terminus as an effect of seasonally variable rheology in the ice mélange; stiffened mélange in winter suppresses major calving events, enabling the terminus to move forward; while in summer, a weaker mélange can no longer prevent major iceberg calving, and the terminus retreats. They used a force balance analysis to demonstrate that large-scale (full-glacier-thickness icebergs) calving events are not likely to occur when the ice front is well grounded. Based on this, they suggested that one of the necessary conditions for frequent full-glacier-thickness iceberg calving at Jakobshavn Isbær is a floating or close-to-floating terminus in summer.

Currently, it is challenging to observe grounding line position directly when it lies near the calving front. However, this can be inferred from observations of ice motion (Henderson and Riedel, 2007; Rignot et al., 2011; Rosenau et al., 2013). For many marine-terminating glaciers, ice speed is affected by ocean tides (e.g., Makinson et al., 2012; Podrasky et al., 2014; Voytenko et al., 2015a). At Jakobshavn Isbær, Podrasky et al. (2014) used GPS and theodolite data obtained in a 2-week campaign in middle to late August 2009 to study velocity response to ocean tidal forcing near the terminus of Jakobshavn Isbær. After removal of a high background speed and perturbations caused by a single calving event, tidal forcing explained a significant fraction of the remaining signal. Based on the fast decay of tidal response upstream, they concluded that the terminus region is very nearly grounded during summer months. Rosenau et al. (2013) used photogrammetric time-lapse imagery to estimate grounding line migration and calving dynamics at Jakobshavn Isbær. They found that the grounding line retreated 3.5 km from 2004 to 2010, with an ephemeral floating tongue during the advance season.

In this study, we use ice velocity and elevation time series observed with terrestrial radar interferometry (TRI) to analyze grounding line position and tidally affected ice flow. Previous work (Peters et al., 2015; Voytenko et al., 2015a, b, c, 2017; Holland et al., 2016; Xie et al., 2016) has shown that TRI can overcome the limitations of GPS (low spatial resolution, difficult to deploy near the calving front), theodolite (low spatial resolution and precision), photogrammetry (low reliability in bad weather and at night), and satellite observations (low temporal resolution). Here we use TRI measurements obtained in three summer campaigns, but at different stages (early versus late summer) of the calving season, to investigate tidal response and the evolving glacier front through Jakobshavn Isbær’s calving season.

2 Data acquisition

We observed the terminus of Jakobshavn Isbær in three summer campaigns in 2012, 2015, and 2016. Each campaign obtained a continuous record of velocity and elevation change over 4 to 13 days. The TRI instrument (Gamma Portable Radar Interferometer) is a real-aperture radar operating at Ku-band (1.74 cm wavelength) and is sensitive to line-of-sight (LOS) displacements of ∼ 1 mm (Werner et al., 2008). It has one transmitting and two receiving antennas, which allows for high spatial and temporal resolution measurements of both displacement and topography. The antennas are rigidly attached to a rack structure, which sits on a motor that rotates around a fixed vertical axis. In 2012, the instrument was deployed on a tripod reinforced with sandbags, with the calving front ∼ 3–6 km away. In 2015 and 2016, the instrument was mounted on a metal pedestal connected to bedrock with 10 cm bolts and protected by a radome to eliminate disturbance from wind and rain, with the calving front ∼ 2–5 km away. In all three campaigns, the radar scanned to a maximum distance of 16.9 km, generating images with both phase and intensity information. The resolution of the range measurement is ∼ 1 m. The azimuth resolution varies linearly with distance and varies as the arc length I = D · A, where D is the distance to the radar and A is the azimuth angle steps were 0.2°, resulting in an azimuth resolution of 7 m at 2 km distance, 14 m at 4 km, etc. Other parameters in these measurements are listed in Table 1. Figure 1 shows the spatial coverage of measurements in each campaign.

3 Data analysis

3.1 TRI data processing

TRI data were processed following Voytenko et al. (2015b): (1) slant range complex images were multi-looked to reduce noise; (2) interferograms were generated between adjacent scans; and (3) a stationary point on rock was chosen as a reference for phase unwrapping. Unwrapped phases were then converted to LOS velocities. We define LOS velocity as positive when ice moves towards the radar and negative when ice moves away from the radar. All results were resampled into
10 m × 10 m pixel spacing maps unless otherwise specified, with a bicubic spline interpolation algorithm. To georeference the TRI results, we used a Landsat-7/8 image acquired during (if not possible, then with a < 2-day time difference) the observation period as a reference. By fixing the radar location and horizontally rotating the intensity image, a rotation angle was estimated based on the best match of distinct surface features (e.g., coast line, ice cliff, icebergs); thus TRI-derived results were georeferenced into the earth reference system. In this study, we use the polar stereographic projection to minimize distortion. Notice that the TRI instrument measures LOS intensity and phase information. Converting LOS data into x–y grid coordinates induces some distortions due to topography, especially in the mélange close to the radar, where the height differences are largest. The radar location in 2012 was ∼280 m above local sea level, and in 2015/2016 ∼200 m above local sea level. A simple calculation based on geometry shows that distortion due to topography is < 15 m. There are two other error sources in georeferencing TRI data: (1) radar position error (it was measured with a single-frequency GPS, with location error estimated at less than 10 m) and (2) rotation error in matching TRI and Landsat images. By comparing georeferenced TRI images with different Landsat-7/8 images, we found no visible mismatch larger than four pixel widths of the satellite images. We thus assess that the coordinate error in georeferenced TRI results is < 60 m, i.e., smaller than four pixels (typically < 2 pixels) of Landsat-7/8 panchromatic images. Moreover, because the radar was deployed on a fixed point in each campaign, and we used the same radar coordinates and rotation angle in georeferencing for each campaign, the error due to georeferencing will not affect our time series analysis. Other errors in TRI data, such as phase variations associated with variable atmospheric water vapor between adjacent scans, are difficult to model but should not be significant in the near field given the 1.5–3 min repeat time. To minimize water vapor effects, we only analyzed data within 10 km of the radar unless otherwise specified.

TRI data obtained in 2015 have been previously discussed in Xie et al. (2016). The same data are used here, but we added 17 h of additional data obtained before the period analyzed by Xie et al. (2016). The additional data were acquired when the instrument was in an experimental mode: rather than 150° of scan, the scanned arc was sometimes set to different values, and the repeat time was sometimes 1 or 2 min rather than 1.5 min. Otherwise, the additional data have the
same quality as subsequent acquisitions. We processed the additional data with the same standards and converted them into the same reference frame as the remaining 2015 data.

Except for several rapid changes in velocity caused by calving events, the processed results from 2015 and 2016 have good continuity. However, velocities from 2012 have some significant offsets (Fig. S1a in the Supplement). Most of these offsets reflect phase unwrapping errors, reflecting incorrect integer multiples of microwave cycles applied during the phase unwrapping process. The repeat time in 2012 (3 min) was longer than the other two years, and ice motion relative to adjacent areas in the radar LOS during that interval could exceed one radar wavelength. We fixed these phase offsets in three steps: (1) estimate the velocity time series at a single point on the ice (with integer multiples of microwave cycles corrected); (2) use this kinematic point as the reference point for phase unwrapping to get relative velocities for all other mapped points; and (3) add the velocity model from step 1 to the relative velocities. We compared this new velocity map with velocities estimated by feature tracking (done with Open Source Computer Vision Library: https://opencv.org/; uncertainty is typically \(<1 \text{ m d}^{-1}\) for a pair of images separated by 1 day), which is independent of interferometry and does not require phase connection. The phase jumps are greatly reduced, and we believe the resulting velocity time series are an accurate indicator of ice motion. Details are given in Sect. S1 and Figs. S1–S5.

3.2 Tidally driven ice motion analysis

The glacier directly interacts with the ocean at the calving front. By changing back pressure on this front, ocean tides are known to influence the behavior of some marine-terminating glaciers (Walters, 1989; Anandakrishnan and Alley, 1997; Podrasky et al., 2014). Besides back pressure, a full-Stokes nonlinear viscoelastic model (Rosier and Gudmundsson, 2016) suggests that, when there is a floating ice tongue, tidal flexural stress can also be an important forcing for marine-terminating glaciers. In addition, tidal variation can influence basal friction at the ice–bed interface, thus changing the sliding rate of the glacier (e.g., Walker et al., 2013; Voytenko et al., 2015a).

For all three campaigns, velocities near the terminus show significant semi-diurnal variation and perhaps a small diurnal signal. Figure 2 shows the power spectral density (PSD) analysis for selected data in 2016. PSDs for 2012 and 2015 are shown in Sect. S2, Figs. S6 and S7. Previous studies indicate that, apart from calving events, short-term ice velocity variations at Jakobshavn Isbræ are well described with simple tidal response models (e.g., Rosenau et al., 2013; Podrasky et al., 2014). Diurnal variation caused by surface melting may also contribute to velocity variation. This has been observed at both Jakobshavn Isbræ (Podrasky et al., 2012) and Helheim Glacier (Davis et al., 2014). Due to the short time span of our data, it is not possible to recover the full temporal spectrum of ice velocity variations. Instead, we focus on the largest spectral components of the velocity field.

There was no tide record in the fjord near the terminus during our campaigns. Podrasky et al. (2014) analyzed a 14-day tide record in the fjord within 5 km of the calving front obtained in August 2009 and compared it with a longer record from Ilulissat. The two datasets show close agreement, with no measurable delay in time, and a maximum difference in stage \(<10 \text{ cm}\). Thus they used the longer record of tides at Ilulissat to analyze the tidal response of the glacier. We also used analyzed tidal constituents from the long-term record at Ilulissat to predict tides in the fjord during our campaigns. Richter et al. (2011) applied harmonic tidal analysis to 5 years of long-term sea-level records at Ilulissat and estimated that the largest three tidal constituents are K1, M2, and S2, with amplitudes of 0.331, 0.671, and 0.273 m, respectively. These three constituents account for \(>95\%\) of all the analyzed tidal constituents. Figure S8 shows the predicted tide and tidal rate (defined as the first time derivative of tidal height) during the 2015 campaign, when we had a mooring deployed at the mouth of the fjord (red hexagon in Fig. 1) that recorded tidal height. There are only small differences between measured tide or tidal rate with predictions using the three largest constituents. In the following analysis, we focused on ice velocities with the same frequencies as the K1, M2, and S2 tide constituents. Other components of tidal motion with similar frequencies will be aligned into these three constituents. For example, diurnal variation caused by surface melting with a period of \(\sim 1 \text{ d}\), if it exists, will not be separable from K1 with a period of 1.0027 d.

Many tidal response models analyze the response of ice position to tidal height variation (e.g., Davis et al., 2014; Podrasky et al., 2014). However, our TRI measurements are only sensitive to LOS displacement. The corresponding velocity derived by interferometry is the first time derivative of LOS displacement. Velocity can be converted to position by integration; however, due to data gaps and the nonlinear behavior of the velocity time series, integration of velocity time series may introduce artifacts. Therefore, we used ice velocity instead of position and analyzed the response of ice velocity to tidal rate. The amplitude of variation is magnified by frequency (signals with higher frequencies will have larger ranges of first time derivative; see Sect. S3), but the phase difference is unchanged by differentiation.

Before the tidal response analysis, we used the modified Z-score method (Iglewicz and Hoaglin, 1993; also see Sect. S1) to remove outliers. We note that TRI-observed ice motion in the mélange is very sensitive to small calving events, while ice on the glacier is less affected. Due to frequent calving events in the 2012 data, we were not able to accurately model the full time series. Instead, we used data obtained from 6 to 10 August when there was only one small calving event (see Fig. S1) for the following analysis. For the 2015 data, there were many small calving events and a large one at the end (Xie et al., 2016), resulting in a noisy time se-
Figure 2. Stacked power spectral density (PSD) estimates of the LOS velocity time series for selected areas in 2016. Three 0.5 km × 0.5 km boxes (a, b, and c) mark the selected areas. PSD plots are normalized, and each black line represents 1 pixel (10 m × 10 m) in the corresponding box. Red line shows mean value. Blue lines mark frequencies of K1, M2, and S2 tide constituents. On map to the left, dashed orange line shows a significant step change of height in the mélange observed in 2016 (see also Fig. 3a).

Midas. We therefore omitted the 2015 mélange from further analysis. For 2016, a step change in ice elevation (dashed orange line in Fig. 2) was observed, separating the mélange into two distinct parts. Downstream from the step change, ice motion is very noisy and difficult to analyze for periodic signals. Upstream from that, ice velocity variation is similar to the glacier. Therefore, we did not do tidal response analysis for the ice mélange downstream from the step change in 2016. Movies S1, S2, and S3 show all major calving or calving-like (collapse of tightly packed mélange) events observed during the three campaigns, and corresponding changes in the mélange.

For both 2012 and 2015 campaigns, ~4 days of data were analyzed, and a second-order polynomial was used to detrend the time series. For the 2016 campaign, ~13 days of data were analyzed. This time series shows significant responses to a few calving-like collapse events (Fig. 3). We used a function composed of a second-order polynomial + 3 pairs of sines and cosines to estimate the response to calving(-like) events and then removed the polynomial. The function is

\[ V_i = a_j + b_j t + c_j t^2 + \sum_{k=1}^{3} d_k \sin(2\pi f_k t_i) + e_k \cos(2\pi f_k t_i), \]

where \( V_i \) is the observed LOS velocity at time \( t_i \), and \( a_j, b_j, \) and \( c_j \) are coefficients of second-order polynomial for the \( j \)th period, where periods are separated by large calving(-like) events. To better estimate the second-order polynomial, periods with data spanning shorter than 1 day are not used. \( d_k \) and \( e_k \) are coefficients of the \( k \)th periodic component, with frequency \( f_k \) among those of K1, M2, and S2 tidal constituents. Response to calving events and tidal constituents with periods > 2 days is largely eliminated with this procedure. Figure 3 gives an example of the observed and detrended time series. Note that data in 2016 span longer times than 2012 and 2015. To save computational time, we converted TRI images into pixel sizes of 30 m × 30 m for a map-wide analysis.

Detrended time series were passed through a median filter to reduce noise. The kernel size is 3, 5, and 5 for data in 2012, 2015, and 2016, equal to a 9, 7.5, and 10 min time window, respectively. All time series were then analyzed using the method of Davis et al. (2014), which estimates the amplitudes and phases of the three periodic components with the same frequencies as the K1, M2, and S2 tidal constituents. This method allows us to distinguish components with close frequencies (in our case, M2 and S2). We also used a least squares fit to an equation with three frequencies of sines and cosines as an alternative method. The two methods fit the time series equally well, with differences that are insignificant compared to noise. Note that we assume constant tidal response for each campaign, whereas in reality tidal response can have temporal variation due to calving and other processes. However, previous work at Jakobshavn Isbræ (Podrasky et al., 2014) and Helheim Glacier (de Juan et al.,...
Figure 3. (a) Ice velocity estimated by feature tracking using a pair of TRI intensity images separated by 1 day in the 2016 campaign. Dashed white line outlines the area with (nearly) stationary points used to define uncertainty of velocity estimate; the rms of velocity estimates (without detrending) by feature tracking within the dashed outline is < 1 m d\(^{-1}\). (b) TRI-observed LOS velocity time series for a single point, marked by white dot in (a). Grey dots show velocities derived from unwrapped phases, red curve shows the model used to remove perturbations caused by calving events, and black dots show detrended time series offset by −8 m d\(^{-1}\). Blue arrows mark large calving or calving-like collapse events. Orange line shows changes of angle between LOS and 2-D ice velocity direction by feature tracking. The LOS velocity variation for a period longer than 1 d is mostly due to changes in background velocity direction.

2010) shows that this variation will not significantly change the phase of tidal response during a period of few weeks.

Figure 4b, d, and f show maps of phase lag (converted to time in hours) from tidal rate to TRI-observed LOS velocity at the M2 tidal frequency, along with a velocity profile for each campaign. Note that, due to the phase character of periodic signals, dark red on the map represents phase values that are close to dark blue. For example, 12.42 h (period of M2) “equals” 0. Note also that the phase lag maps only show pixels with signal-to-noise ratio (SNR) > 1.5, where we define SNR as

\[
\text{SNR} = \frac{\sigma_{\text{signal}}^2}{\sigma_{\text{noise}}^2}.
\]

We use the root mean square (rms) of the detrended velocity time series to represent \(\sigma_{\text{signal}}\), and rms of the residuals to represent \(\sigma_{\text{noise}}\). We use the M2 tidal signal to illustrate tidal responses since this is the largest tidal constituent. Phase lag maps for K1 and S2 are shown in Fig. S9, with patterns that are similar to M2.
Figure 4. Phase lag map and velocity time series for a profile in each campaign. Grey dots (a, c, e) show detrended LOS velocity time series for a profile along the ice flow line, marked by white dots on the map to the right. Red curve shows best model fit. LOS velocities are offset for clarity. Cyan curve shows tidal rate. Phase lag map (b, d, f) shows M2 frequency signal. Areas where SNR < 1.5 are omitted. Phase lags are converted to times (in hours). In (f), dashed red line shows TRI-derived location of glacier front on 13 June 2016. B1, F1, and F2 mark selected points showing velocity time series in Fig. 6. Note that the amplitude of detrended LOS velocity depends on a number of factors, including tidal response, ice flow direction relative to radar LOS, distance up-glacier, whether the scanned area is glacier or mélange, and (within the mélange) whether the imaged pixel is close to or far from the calving front.
Figure 5. Annual maximum and minimum extents of Jakobshavn Isbræ’s calving front from 2012 to 2016. Solid lines show the glacier front when glacier extent is maximum, and dashed lines when glacier extent is minimum. Glacier front locations are derived from available Landsat-7/8 and Sentinel-2 images in USGS archive. Legends are dates of image acquisition. Lines with triangles, stars, and circles show glacier front locations during TRI campaigns in 2012 (6 August), 2015 (9 June), and 2016 (13 June), respectively. Background for this figure is a Landsat-8 image acquired on 4 June 2015.

Figure 4 shows two types of phase lag patterns. For 2012, LOS velocity of ice in the mélange has ~0 phase lag to tidal rate, whereas the phase lag increases sharply at the calving face, to ~8.5 h on the glacier front. For both 2015 and 2016, there is a narrow zone at the glacier front that is in phase with the tidal rate, with phase lag close to 0. Upstream from that, phase lag increases to ~8 h.

4 Discussion

4.1 Grounding line variation in a calving season

One hypothesis concerning the annual cycle of advance and retreat of Jakobshavn Isbræ is that a floating tongue grows in winter and disappears in late summer (Joughin et al., 2008; Amundson et al., 2010). However, there are no direct observations through a full calving season. We addressed this by assuming consistent behavior over the 5-year observation period and considering our data to be a representative sample of the good inverse correlation between seasonally varying speed and length of ice tongue (Joughin et al., 2008, 2014).

Rosenau et al. (2013) looked at the cross-correlation coefficient between tidal height and the vertical component of ice trajectory to estimate grounding line migration. This approach assumes that the only force that drives vertical ice motion is tide rise and fall. From an analysis of optical images, they found no evidence of floating in mid-July 2007 (~6-day duration), a ~500 m wide floating zone from 8 to 9 August 2004 (~1-day duration), and an even wider floating zone from late spring to early summer 2010 (~29-day duration). Podrasky et al. (2014) applied a tidal admittance model to analyze both horizontal and vertical responses to tidal forcing at Jakobshavn Isbræ. They found rapid decay of admittance at the glacier front, corresponding to small (~2 and ~0.7 km for horizontal and vertical, respectively) e-folding lengths (the distance over which the amplitude decreases by a factor of e), concluding that the glacier front was very nearly grounded in late August 2009.

TRI-derived LOS velocities reflect several forcings. Surface meltwater-induced velocity variation is a quasi-diurnal signal. Podrasky et al. (2012) detected an amplitude of up to 0.1 m d^{-1} diurnal signal 20–50 km upstream from the terminus of Jakobshavn Isbræ. The timing of the diurnal maxima was ~6 h after local noon, consistent with surface melting. Within 4 km of the glacier front, Podrasky et al. (2014) found diurnal variations that are 0.5–1 times the amplitude of tidally forced variations, with a maxima 10.9–11.7 h after local noon. At Helheim Glacier, Davis et al. (2014) identified a signal with peak-to-peak variation of ~0.7 m d^{-1} in glacier flow speed at a site close to the terminus, likely associated with changes in bed lubrication due to surface melting. While surface meltwater can cause a diurnal component in ice velocity, it should have no direct influence on semi-diurnal signals, which are the dominant signals observed in all three of our campaigns. Supraglacial lake drainage events could be another possible forcing process, though they were not observed near the terminus during our campaigns. Upstream from the terminus, supraglacial lake drainage events occur but are sporadic. Podrasky et al. (2012) observed at most three supraglacial lake drainage events near the terminus during three summers from 2006 to 2008. If such events occurred during our data collection periods, the responses are likely to have been eliminated by the detrending process.

The LOS velocity variation contains two components of ice motion: (1) vertical motion and (2) horizontal motion. For all three campaigns, the radar was always located higher than the ice surface in the mélange and the first 3 km of the glacier. In this case, the TRI-observed LOS velocity component is:

\[ V_{los} = \frac{1}{\sqrt{\left(\frac{1}{\sigma_h^2 + \sigma_v^2}\right)^2 + 1}} \frac{dh}{dt}, \]
where \( L \) is the horizontal distance between the radar and target, \( H_0 \) is the mean height different between the radar and target, \( h \) is the vertical movement relative to \( H_0 \), and \( \frac{dh}{dt} \) is the vertical component of ice velocity (see geometry in Fig. S10). We assume that, for floating ice, \( \frac{dh}{dt} \) is correlated with the tidal rate. Hence \( \frac{dh}{dt} \approx \) tidal rate in the mélange, and less than that for the glacier, but it can be close if ice near the glacier front is very weak, similar to what Voytenko et al. (2015a) found at the terminus of Helheim Glacier. For grounded ice, \( \frac{dh}{dt} \) variation should have a much smaller amplitude than tidal rate variation. Horizontally, for all three campaigns, ice on almost the entire glacier moves towards the radar (LOS velocity is positive; see Figs. S3, S4, and S5). Previous studies suggest that several mechanisms are acting simultaneously, and there is no single defined phase relation between tide variation and ice speed (e.g., Thomas, 2007; Abalgeirsdóttir et al., 2008; Davis et al., 2014; Podrasky et al., 2014). However, at the terminus of Jakobshavn Isbær, Podrasky et al. (2014) found that glacier speed and tidal height are anti-correlated. This likely reflects variation of back-pressure forcing associated with tide rise and fall. We have not attempted to derive a comprehensive model for ice velocity variation caused by changes of back pressure or other factors. Instead, we adopt the admittance parameters estimated by Podrasky et al. (2014) to assess a near-upper bound of along-flow-line velocity variation. Using theodolite and GPS observations near the ice front, Podrasky et al. (2014) estimated horizontal and vertical tidal admittances of < 0.12 and < 0.15, respectively. In terms of phase, tide-induced vertical motion is in phase with the ocean tide, while horizontal velocity is anti-correlated with tidal height; i.e., horizontal velocity maxima are concurrent with the inflection points of tidal rate. By assuming the glacier was under the same conditions as the time when Podrasky et al. (2014) did their measurements, we predict ice velocities near the glacier front. In Fig. 6a, F1 and F2 correspond to the two points marked with purple triangles in Fig. 4f. For each point, two components of ice velocity were predicted and projected onto the LOS direction to the radar: (1) vertical velocity by using tidal admittance of 0.15, and time lag of 0 to tidal rate, shown by solid black curve, and (2) horizontal velocity by using tidal admittance of 0.12, and anti-correlated with tidal height, shown by the dashed black curve. The red curve shows the sum of these two components. Podrasky et al. (2014) inferred that the glacier front was very nearly grounded during their observation period, and both horizontal and vertical tidal admittances dropped dramatically upstream. While we use the upper bound of the tidal admittance by Podrasky et al. (2014), the amplitudes of our predicted velocities are almost the maxima for grounded ice. However, as shown in Fig. 6a, predicted tide-induced vertical velocities have far smaller magnitude than our TRI-derived velocities – the horizontal component is larger but is negatively correlated with TRI observations. Therefore, we reject the hypothesis that ice near the glacier front in early summer 2016 was nearly grounded as during the observation period of Podrasky et al. (2014) in late summer. For comparison, we also plot predicted LOS velocities by assuming ice was in a free-flotation state, shown in blue. This is in phase with the TRI derived velocities, although the magnitude does not fully explain the larger signals observed by TRI. Possible reasons are discussed below. Ice located in the low-phase-lag zone (dark red or blue in Fig. 4d) in 2015 yields similar results. For ice further upstream in 2015 and 2016, and almost the entire glacier front of 2012, we cannot reject the possibility of a near-grounded
basal condition, because the admittances by Podrasky et al. (2014) can then produce LOS velocities that are sufficiently large and correlated with TRI observations. Figure 6b shows predicted (red curve) and observed (grey dots) velocity of a surface point (B1 in Fig. 4b) that is immediately adjacent to the glacier front during our 2012 campaign. They have similar amplitude and phase, though the maxima of TRI-observed velocity are not exactly concurrent with the inflection points of tidal rate. Instead, they are slightly earlier (∼0.5 h) than the inflection points. We presume that ice in the high-phase-lag zone in Fig. 4 is either grounded or nearly grounded.

Based on this analysis, we hypothesize that during early summer 2015 and 2016 there was a narrow zone of floating ice near the glacier front, which is at least the width of the low-phase-lag zone (∼1 km). However, we are unable to determine if ice more than 1 km from the glacier front is grounded or not. The annual maximum and minimum extents of the ice front (solid/dashed lines in Fig. 5) support this hypothesis: the low-phase-lag zone on the glacier during both the 2015 and 2016 observations coincides with the transition zone between maximum and minimum glacier front. In contrast, for the 2012 data, the glacier front was close to the annual minimum. Additional evidence to support this hypothesis comes from the ice surface elevation map. Figure 7 shows the median average DEM from estimates of a 1-day TRI measurements for each campaign. In 2012, near the centerline of the main trunk, surface ice elevation increases dramatically near the glacier front, to >120 m in <1 km distance from the glacier front. In contrast, in 2015 and 2016, ice elevation increases more slowly, with a ∼1 km wide zone that is <110 m higher than local sea level. In this low-elevation zone, overall buoyancy could make conditions favorable for a floating glacier front during early summer (2015 and 2016 data).

During the time span of our TRI campaigns, the glacier front maintained a relatively constant position, with ∼3 km ice advance and retreat per year. Time series of satellite images also suggest that in late summer to early autumn the glacier front usually stabilizes near the minimum position for a few weeks before a steady advance. Using the TRI campaign in 2012 as a proxy for late-summer conditions, and campaigns in 2015 and 2016 as proxies for early-summer conditions, we infer that from 2012 to 2016 Jakobshavn Isbrae had a floating tongue in the early stage of the calving season. Undercutting and tidal flexure then weakened the floating ice, leading to large calving events in subsequent months. During the calving season, calved ice surpassed ice flow into the terminus zone, causing the glacier front to retreat. In late stages of the calving season, the glacier had lost the majority of its floating tongue, and the ice front became grounded or nearly grounded.

4.2 Other sources of forcing

Figure 6a shows that, even when assuming ice is free-floating near the glacier front, LOS velocity variation generated by tide rise and fall is insufficient to explain the observed velocity time series. Ice velocity variation caused by surface melting, if in phase with tidal rate, can increase the overall velocity variation. In this study, we did not separate the quasi-diurnal signal associated with surface melting from similar tidal components. However, there is some evidence of such a signal. As shown in the normalized PSD in Fig. 2c, the diurnal constituent is less obvious than in Fig. 2a and b: assuming speed maxima caused by surface melting lags local noon by 6 h, it will be in phase with the K1 ocean tide rate. Due to the geometry difference, TRI-observed LOS diurnal tidal signal will be superimposed on a negative (box C) or positive (box A and B) diurnal signal associated with surface melting, decreasing or enhancing the observed signal. Thus the diurnal constituent in Fig. 2c is smaller than the other two areas. However, surface melting should not make a significant contribution to semi-diurnal signals, as it is a diurnal phenomena. In addition, most sources of forcing would induce longitudinal velocity variations, and their signals should attenuate significantly near the glacier front due to the LOS geometry. The large additional variation shown in Fig. 6a has a significant component that is not parallel to long-term ice flow motion, i.e., in the cross-flow-line direction; thus it cannot be mainly caused by surface melting. We therefore studied points moving in a near-perpendicular direction to LOS, where along-flow-line motion (e.g., velocity variation due to surface melting) is likely to be negligible in the TRI data. The 2016 data are appropriate for this study.

We focused on three points in the mélange (Fig. 8a). The velocity estimates from both interferometry and feature tracking suggest that their along-flow line velocities are almost perpendicular to the radar LOS direction (within ±5 of 90°). Any longitudinal variation would be trivial when projected onto the LOS direction. Figure 8b shows that the LOS velocity variation caused by up-and-down ice motion that is directly related to tides can only explain about half of the observed signal. The extra signal has a strong correlation with tidal rate, with an amplitude of ∼1 m d⁻¹ (∼0.1 m in displacement). This phase relation suggests that either bed topography or tidal currents are responsible for the signal that is non-parallel to long-term ice flow motion. Bed topography is not likely to be the main contributor, as it is more likely to affect glacier motion rather than mélange motion, unless mélange ice is strongly attached to the glacier. There is no ocean current record during our campaigns near the glacier front, and available models for the ice fjord are too coarse. However, as Doake et al. (2002) have discussed, the usually accepted drag coefficient between ice and water is not likely to create enough force to drive ice motion to a sufficient magnitude. To fully explain the periodic non-parallel signal, we need to either assume a very rough surface for ice below the
Figure 7. DEM for the glacier front, derived from median average of DEM estimates separated by 2 min during a 1-day period. For each subplot, red dot shows location of the radar, and pink contours show bed bathymetry in meters (An et al., 2017). Dashed red line shows the glacier front from TRI image. Note that in 2016 it was not possible to distinguish a portion of the glacier front from TRI measurements; hence it is not marked on the map. The background image for (a) was acquired on 6 August 2012 by Landsat-7; white stripes are data gaps. Background image in (b) was acquired on 4 June 2015 by Landsat-8. Background image in (c) was acquired on 13 June 2016 by Landsat-8. Note that uncertainty increases with distance to the radar. Mean elevation of the black box (1 km × 1 km outlines in a–c) immediately adjacent to the glacier front is 99, 109, and 124 m for 2012, 2015, and 2016, respectively; we use this to represent the mean elevation within 1 km to the glacier front. Mean elevation of the black box upstream (1 km × 1 km outlines in b and c, 2–3 km to the glacier front) is 154 and 158 m for 2015 and 2016, respectively. Black, blue, and red line in (d) show elevation profiles along a transect marked (grey lines in a–c). These transects have the same location in space. In (e), the distance of each transect is normalized so that the glacier fronts are in the same position.

5 Conclusions

High spatial and temporal resolution measurements of the time-varying velocity field at the terminus of Jakobshavn Isbræ were acquired with terrestrial radar interferometry. Ocean tides modulate glacier velocity, and this modulation can be used to infer the location of grounding line. The phase relation between ice velocity and tidal rate suggests a ∼ 1 km
Figure 8. (a, b) TRI observed ice motion that is non-parallel to long-term ice flow motion in the mélange of 2016. In (a), color map shows LOS velocity by interferometry, from a 1-day median average. Arrows show velocity estimates from feature tracking projected onto the LOS direction (dark red when ice moves towards the radar and dark blue when ice moves away). Dashed grey line shows glacier front location from TRI image. Black square, blue triangle, and red star mark three points where 2-D velocity direction is nearly perpendicular to radar LOS. Their LOS velocity time series are shown in (b). Note that the point with blue triangle marker corresponds to the marked point in Fig. 3a. Top three rows in (b) show TRI-observed LOS velocities for selected points; cyan curves are predicted LOS velocities based on the imaging geometry, assuming ice is free-floating. LOS velocities are offset for clarity. Bottom row shows residual time series by subtracting the cyan curves. (c, d) TRI observed ice motion that is non-parallel to long-term ice flow motion on the glacier front for 2015. Colors and arrows in (c) represent the same parameters as in (a). A point immediately adjacent to the glacier front was chosen, marked by black square, with its LOS velocity observed with TRI and predicted by tide variations shown in (d). Cyan curve in (d) shows predicted LOS velocities.

wide floating zone in early summer of 2015 and 2016, where TRI-observed velocity variation contains ice up-and-down motion caused by tide rise and fall, and perhaps a component that is non-parallel to long-term ice flow motion due to tidal currents. The floating zone moves together with calved ice through most of the calving season. However, in late summer 2012, there is no evidence of a floating ice tongue. We hypothesize that Jakobshavn Isbræ maintains a short floating tongue from winter to early summer, when ice flow exceeds ice loss by calving and the glacier front advances. In summer, iceberg calving surpasses ice flow, and the glacier front retreats, becoming nearly grounded by late summer. TRI-derived digital elevation models support this hypothesis: in early summer, there is a ~1 km wide zone with relatively thin ice (<110 m) above local sea level; in late summer, ice thickness near the glacier front increases dramatically and buoyancy is insufficient to support a floating glacier front.
Data availability. Landsat-7/8 and Sentinel-2 images were downloaded through the U.S. Department of the Interior U.S. Geological Survey (2018).

The Supplement related to this article is available online at https://doi.org/10.5194/tc-12-1387-2018-supplement.

Competing interests. The authors declare that they have no conflict of interest.

Acknowledgements. We acknowledge Denise Holland at the Center for Global Sea Level Change at New York University Abu Dhabi for organizing the field logistics for the 2015 and 2016 campaign. Judy McInrath of the University of South Florida is thanked for help in the 2012 fieldwork. This research was partially supported by NASA grant NNX12AK29G to Timothy H. Dixon. David M. Holland acknowledges support from NYU Abu Dhabi grant G1204, NSF award ARC-1304137, and NASA Oceans Melting Greenland NNX15AD55G. Surui Xie thanks Nicholas Voss at the NYU Abu Dhabi grant for Global Sea Level Change at New York University Abu Dhabi to Timothy H. Dixon. David Holland thanks William Roach and Judy McIlrath of the University of South Florida for helpful discussions. Comments from editor Olaf Eisen and two anonymous reviewers are greatly appreciated.

Edited by: Olaf Eisen Reviewed by: two anonymous referees

References


Motyka, R. J., Truffer, M., Fahnestock, M., Mortenson, J., Rysgaard, S., and Howat, I.: Submarine melting of the 1985 Jakobshavn Isbrae floating tongue and the trigger-


Appendix III: Licence and reprint of *Xie et al., 2019, NC*

Open access

*Nature Communications* is an open access journal.

As of January 2016, the journal only publishes open access content, and legacy subscription content has been made freely accessible alongside the open access articles published in *Nature Communications* prior to 2016.

Creative Commons Licenses

*Nature Communications* articles are published open access under a CC BY license (Creative Commons Attribution 4.0 International License). The CC BY license allows for maximum dissemination and re-use of open access materials and is preferred by many research funding bodies. Under this license users are free to share (copy, distribute and transmit) and remix (adapt) the contribution including for commercial purposes, providing they attribute the contribution in the manner specified by the author or licensor (read full legal code).

Some historical papers have been published under a non-commercial license. Users may request permission to use the works for commercial purposes or to create derivative works by emailing permissions@nature.com.

Under Creative Commons, authors retain copyright in their articles.

Visit our open research site for more information about Creative Commons licensing.

Benefits of open access

Open access publication can lead to:

- Increased citation and usage
- Greater public engagement
- Faster impact
- Broader collaboration
- Increased interdisciplinary interaction

For more information about the benefits of open access, please visit our open research site.

https://www.nature.com/ncomms/about/open-access
Compliance with funder mandates

Nature Research’s services and policies ensure that authors can fully comply with the public access requirements of major funding bodies worldwide. For information and advice about compliance, consult our open access policy information, or contact us.

Article processing charges

For more information about our Article processing charges (APC) and applications for discretionary APC waivers please visit our page on Article processing charges.

Nature Communications

ISSN 2041-1723 (online)

© 2020 Springer Nature Limited
Rapid iceberg calving following removal of tightly packed pro-glacial mélange

Surui Xie 1, Timothy H. Dixon 1, David M. Holland 2,3, Denis Voytenko 1 & Irena Vaňková 2,3,4

Iceberg calving is a major contributor to Greenland’s ice mass loss. Pro-glacial mélange (a mixture of sea ice, icebergs, and snow) may be tightly packed in the long, narrow fjords that front many marine-terminating glaciers and can reduce calving by buttressing. However, data limitations have hampered a quantitative understanding. We develop a new radar-based approach to estimate time-varying elevations near the mélange-glacier interface, generating a factor of three or more improvement in elevation precision. We apply the technique to Jakobshavn Isbræ, Greenland’s major outlet glacier. Over a one-month period in early summer 2016, the glacier experienced essentially no calving, and was buttressed by an unusually thick mélange wedge that increased in thickness towards the glacier front. The extent and thickness of the wedge gradually decreased, with large-scale calving starting once the mélange mass within 7 km of the glacier front had decreased by >40%.
Previous work suggests that increasing ice discharge in marginal areas at or close to the glacier front is a major process contributing to recent ice loss in Greenland. Mass loss rates from marine-terminating glaciers are generally more variable than those from other glaciers because of the influence of the time-varying ice–ocean interface. Several factors affect ice discharge rate here, including ocean water temperature, time-varying water levels, terminus position, and mélange extent and strength. Better understanding of ice dynamics at the termini of marine-terminating glaciers has the potential to reduce uncertainty in total mass balance estimates of Greenland and improve projections of future sea-level change. However, direct observations are challenging. Here, we develop a new approach to derive precise glacier and mélange surface elevation maps with high temporal resolution (2-minute interval) over a broad region using a terrestrial radar interferometer (TRI). We apply this approach to the terminus of Jakobshavn Isbiser, a major Greenlandic glacier with persistent proglacial mélange. Jakobshavn Isbiser, Greenland’s fastest moving glacier, has retreated tens of kilometers in the last few decades. Increased subsurface melting triggered by incursion of warm ocean water has been suggested as an important contributor. The glacier’s terminus is now embedded in the ice sheet, with a relatively steady position, despite some seasonal advance and retreat. However, it is unclear how stable the present terminus position will be in the longer term, since Jakobshavn Isbiser has a retrograde bed. A previous study suggests that this type of glacier is conditionally stable, with stability affected by the buttressing effect of an ice-shelf. Other work has shown that mélange in front of Jakobshavn Isbiser can be characterized as a weak granular ice sheet that transmits stress from the fjord back to the glacier terminus, and the buttressing force (lateral load) can be large enough to inhibit the initiation of large-scale calving events. It has also been suggested that the buttressing force on the glacier terminus depends on the thickness of the mélange.

To better understand the influence of mélange on calving, we analyzed time series of digital elevation models (DEMs) derived from TRI observations of the terminus of Jakobshavn Isbiser, allowing us to monitor changes in mélange thickness. Ice flow and glacier calving were also analyzed using TRI and satellite data. Our results reveal the details of mélange behavior during a period of glacier quiescence, providing evidence that tightly-packed mélange can suppress iceberg calving.

**Results**

**TRI mapped elevation time series.** We measured time-varying elevations of the terminus of Jakobshavn Isbiser with a TRI during a ~13-day campaign from 7 to 20 June 2016 (Fig. 1a). We focus on the main (southern) branch of the glacier within 10 km of the radar. Glacier ice here is significantly thicker and moves faster than ice in the northern branch. We used 2-minute intervals between scans. A new approach was developed to improve accuracy and precision of the height estimates. We used a high precision DEM (ArcticDEM, from the Polar Geospatial Center, University of Minnesota: https://www.pgc.umn.edu/data/arcticdem/) for stationary rock areas to minimize TRI errors (Methods). Errors due to systematic bias in the ground reference point and radar geometry were corrected using a priori ground elevations from ArcticDEM. Jumps in the height estimate due to phase unwrapping errors were corrected based on their relation to phase jumps. Elevation estimates are relative to a flat surface defined by fjord water using 2% of measured mélange surface heights within a polygon area that has few large icebergs (see Methods and Supplementary Fig. 1). Accuracy and precision of the DEM time series were assessed by computing root-mean-square deviations of time series for representative stationary rock points and slow-moving ice points, and comparison to predicted tides. The derived time series based on median filtering (30-minute time window) have a height uncertainty of ~20 cm at 3 km and ~70 cm at 6 km from the radar. Height uncertainty increases with the square of line-of-sight (LOS) distance to the radar (see Methods and Supplementary Figs. 1–3). This method provides more than a factor of three precision improvement compared with previous approaches, allowing resolution of new processes within the proglacial mélange, such as melting, collapse, and tide-induced elevation changes.

Figure 2a shows a 1-day median DEM. The elevation time series for representative points in the mélange (c–e) and on the glacier (f) are shown in Fig. 2c–e. Except for perturbations caused by calving-like collapse events within the mélange (melange collapses that are similar in some respects to iceberg calving, see Supplementary Movie 1), tidally induced surface elevation changes in the mélange are well-resolved. Elevation profiles and inferred thicknesses (Fig. 2b, Supplementary Figs. 4 and 5) show a distinct step-like change in surface elevation of ~10 m located 2–4 km from the glacier front. The thick mélange upstream from the elevation step-change has a wedge-like shape, thickest at the glacier front, tapering downstream. Inferred thickness of the mélange (based on TRI-derived surface elevations and assuming hydrostatic equilibrium) near the glacier terminus exceeds 400 m (Supplementary Fig. 4). During our 2-week observation period, the elevation step-change migrated toward the glacier via several calving-like collapse events, progressively removing the downstream edge of the mélange wedge (Supplementary Movies 1 and 2). By the end of our campaign, the elevation step-change in the mélange was ~2 km from the glacier front (Fig. 3, Supplementary Movies 1 and 2).

Tightly packed mélange wedge suppressed calving. Satellite images show that the main trunk of the glacier did not calve for...
calving events as those with block size >0.25 km² and causing significant mélange motion; minor calving events are those in which visible blocks calved, but the proglacial mélange remains largely unchanged. Satellite observations show that large-scale calving events resumed within 1.5 days after the end of our campaign, causing ~9 km² ice loss from the glacier front within 8.5 days (Fig. 3–f). Previous studies have suggested that mélange strength and iceberg calving rate (defined here as calved ice mass per day) are inversely related10,11. We hypothesize that calving was suppressed by the buttressing force from tightly packed proglacial mélange during a ~30-day period, from ~17 days before the beginning of our campaign until its end. Large-scale calving occurred once the mélange weakened sufficiently (i.e., the elevation step-change in the mélange migrated to <2 km from the glacier front within 1.5 days after the end of our TRI campaign). Our new DEM time series allow some aspects of this process to be quantified for the first time.

Mélange ice mass loss. The buttressing force of the mélange is positively correlated with sea ice and/or iceberg thickness and concentration10–11. Mélange ice mass (or thickness) may, therefore, be a useful proxy for mélange strength. To estimate mélange strength changes, we use total ice mass defined within a fixed proglacial Lagrangian area, and investigate changes in this mass. The close match between tidal and mélange heights (Fig. 2c–e) implies that the mélange is near hydrostatic equilibrium. A bed elevation map (contour in Fig. 2a) indicates that the fjord depth here is larger than the mélange thickness. Archimedes' principle thus allows us to use our elevation time series to estimate temporal changes in mélange thickness and mass.

Several mechanisms control loss or gain of mélange ice within a given region: first, downstream advection and divergence of ice driven by glacier motion and outflow of fjord water; second, gravity-driven collapses in over-thickened mélange that enhance advection of mélange near the elevation step-changes; third, melting of the mélange driven by contact with warm air and water. Our DEM time series allow us to separate changes caused by some of these mechanisms. We calculate melt thinning (overall thickness decrease) rate based on changes of surface elevation, which are also affected by mélange divergence (details below and in Methods). To separate divergent thinning from melting, we select a box (dashed rectangle R in Fig. 5a) within 2 km of the glacier front, and track it in a Lagrangian reference frame. This selected area remained upstream from the elevation step-change until the end of TRI observations, and exhibited insignificant changes in shape and iceberg distribution pattern throughout the observation period (Fig. 5a-c and Supplementary Movie 1). Thus iceberg fragmentation should have minimal effect on melt thinning rate estimates. Each pixel within the selected area is treated as a cell with independent mobility, and the evolution of the shape and location of the selected Lagrangian area is estimated by feature-tracking24 (Fig. 6 and Supplementary Fig. 6). Mean divergent thinning is determined by area changes of the convex hull (envelope) of all cells. Subtracting divergent thinning from the total thinning yields thinning due to melting (Methods). Because of the density difference between water and ice, TRI-measured changes in mélange surface elevation represent about one tenth of the total mélange thickness reduction. In the selected box R, TRI-derived elevations have a height uncertainty of 0.2–0.3 m, whereas the mélange thinning rate here is 0.8–1.9 m d⁻¹ (Fig. 5m), corresponding to ~0.1–0.2 m d⁻¹ in surface elevation change. Therefore, to allow sufficient signal-to-noise ratio, TRI measurements separated by multiple days are used to estimate the thinning rate. However, if a pair of TRI images is separated by too long a period of time, feature tracking...
correlation decreases and the divergence uncertainty will be larger. We thus estimate average total/divergent/melt thinning rates for 6-day periods. This allows us to measure surface elevation changes that are more than a factor of 2 larger than the uncertainty in the TRI-derived DEMs, and can also give a first order estimate of the reliability of thinning rate estimates by comparison of results from different periods. Figure 5i–l show examples of elevation changes along the major axes of four selected icebergs. Figure 5m–o show the thinning rate estimates. While there is a wide range, the divergent thinning rate is always positive, implying overall extensional motion of the mélange wedge. This may be explained by fjord geometry—the fjord widens with increasing distance from the glacier terminus. Velocity and displacement maps also show that ice motion within the mélange wedge is affected by curvature of the fjord wall (Fig. 6). Figure 5n indicates that the divergent thinning rate increased with time during the observation period, suggesting an overall increase in mélange mobility. We calculated weighted mean total and melt thinning rates and corresponding uncertainties (using one weighted standard deviation), yielding an average
total thinning rate of 1.4 ± 0.4 m d\(^{-1}\), and an average melt thinning rate of 1.0 ± 0.5 m d\(^{-1}\) during the TRI observation period.

Note that we assume a simple buoyancy relation in the above calculations: ice and water have constant densities and melting is treated as an incompressible continuum (so that there is a fixed ratio between melt thickness and above-water height). Allowing ice density to change within a plausible range (917–1000 kg m\(^{-3}\)), and water density to vary within 1027 ± 5 kg m\(^{-3}\) will change the mean total thinning and melt thinning rates by less than 25%. Iceberg shapes can be complex, but are impossible to define from our observations. Previous studies suggest that submerged iceberg shapes can be reasonably approximated by cylinders\(^{25,26}\). Since we only attempt to estimate an average melt thinning rate over an area, errors induced by simplifying icebergs as cylinders should be minor. Overall, these assumptions do not change the trend of our thinning rate estimates.

Our range of melt rate estimates is comparable to estimates of 0.7–3.9 m d\(^{-1}\) surface melt rates measured in summer 2008 at three nearby glaciers in the Disko Bay area\(^{22}\), but is considerably larger than the −0.3 m d\(^{-1}\) estimate for an area further downstream from Jakobshavn Isbrae between 2011–2015 from high-resolution satellite observations\(^{26}\). Our higher melt rate estimate may reflect the different locations of the study areas relative to the glacier front. The selected Lagrangian box R in our study is much closer to the glacier terminus (<2 km) than the study giving ~0.3 m d\(^{-1}\) melt rate further downstream the fjord\(^{26}\).

High subglacial freshwater discharge in summer can enhance melting near the glacier terminus by thermal plume convection, based on observations\(^{28}\) and modeling\(^{29}\) showing that tidewater glacier termini can have very high melt rates (exceeding 10 m d\(^{-1}\) in extreme cases), driven by subglacial freshwater discharge. Other factors, such as inter-annual variabilities in surface air temperatures or water temperatures can also cause differences in ice melt rate estimates.

Assuming that the average melt thinning rate (1.0 ± 0.5 m d\(^{-1}\)) within the Lagrangian box R (Fig. 5a) is representative of the average melt thinning rate of mélange within 7 km of the terminus during the observation period, ice loss from melting accounts for ~40 ± 20% of the observed total decrease of mélange mass in a test Lagrangian area (dashed polygon P in Fig. 7a–c). The rest can be attributed to calving-like collapse events or divergent motion within the mélange that helps to advect ice away. The two processes are not independent. Melting can break the gravity-buoyancy equilibrium, causing calving-like collapses within the mélange. Collapse events can change fjord water stratification and circulation, allowing greater ice mobility.

Figure 7a–c shows selected mélange elevation changes through part of the observation period. Supplementary Movie 2 shows changes during the entire observation period. The elevation step-change in the mélange migrates toward the glacier front (Fig. 7a–d) with significant mélange ice removed by calving-like collapses. Migration of the elevation step-change is not linear: It moves downstream between two calving-like collapses, but jumps upstream during each calving-like collapse (Supplementary Fig. 5c). The TRI data can be used to calculate the change of total ice mass within the test Lagrangian area (dashed polygon P in Fig. 7a–c). By inspecting changes in the TRI and available Landsat-8 and Sentinel-1 images, we can also estimate the glacier’s calving rate in daily increments over ~40 days bracketing the TRI campaign (red line in Fig. 7e). There is essentially no calving from 21 May to 20 June 2016, except for a minor event on 10 June that did not significantly affect nearby mélange (Fig. 3).

The coincidence of a thick, tightly packed pro-glacial mélange wedge and the absence of major calving events during an unusually long period (21 May–20 June; Fig. 4) suggests that tightly packed mélange suppressed calving. Subsequently, mélange melting and removal by calving-like collapses (totaling 1.0 ± 0.1 Gt between 7 and 20 June 2016) reduced the buttressing force, eventually leading to major calving by June 21 or 22 (large-scale calving events occurred on these days, based on inspection of TRI and satellite images, see Fig. 3). Assuming that the average ice thickness at the glacier front is 800 m, then the total ice mass calved over the 8.5 day period from June 20–29, 2016 would be 6.7 ± 0.8 Gt, nearly 3% of Greenland’s average annual mass loss between 2003 and 2010\(^{30}\).

Decrease of mélange buttressing force. A study of Store Gletscher used a longitudinal coupling model to explain speed increase at the glacier front associated with clearing of the mélange, suggesting an inverse relation between ice speed and buttressing force\(^{31}\). In our study, ice speed at the glacier front did not show a significant response to mélange changes during the TRI observation period (Fig. 8b, c1, c2 and Supplementary Fig. 7). However, mélange immediately upstream from the elevation step-change did show rapid increases in speed in response to

---

**Fig. 4** Calving events inferred from TRI and satellite images. Due to limited temporal sampling of the data, we are not able to determine the exact time of each calving event. Instead, we mark each calving event in time defined by the closest two usable images (colored dots), defined as no dense cloud coverage at the glacier front in the satellite images. Black dots at 9 and 21 May 2016 represent the acquisition times of two Sentinel-1 images. Black dots at 10 and 20 June 2016 represent acquisition times of TRI images. Two vertical dotted lines mark the 30-day period between 21 May to 20 June. For each year, the time between the grey dot on the left (first acquisition of satellite image between the study period of the year) and the usable image represent the earliest period without calving. If the image represented by the first blue or light blue dot shows no calving events compared to an image before 00:00 1 March in the corresponding year, we put the grey dot at 00:00 1 March. E.g., Landsat-8 images acquired at 14:54 26 February 2017 to until 14:54 15 April 2017 show no calving event during this period. If there is no grey dot on the left of the corresponding year (in 2008, 2010, 2012, due to cloud coverage on Landsat-7 optical images), then the first light blue dot represents the earliest usable image in the corresponding year. Similarly, grey dot on the right for each year indicates the last usable image that shows no calving event compared to the image shown by the last blue or light blue dot in the corresponding year. If there is no grey dot on the right of the corresponding year, it is either because the last blue or light blue dot marks the last usable image, or some calving events occurred after the last shown date in the corresponding year.

---

NATURE COMMUNICATIONS | https://doi.org/10.1038/s41467-019-10908-4 | www.nature.com/naturecommunications
Fig. 5 Mélange thinning. a–c TRI amplitude images, corresponding to times shown in d–f. In a, dashed rectangle R outlines an initial box used to estimate mélange thinning, convex hulls of the evolving box are shown in d–f; red lines are selected profiles to show examples of mélange thinning in i–l, their evolving shapes and locations are shown in b and c. d–f Tidal height subtracted surface elevation of mélange within the selected box, shapes and locations in e and f are estimated based on feature tracking. g Thinning rate from d to e. h Thinning rate from e to f. i–l Surface elevation profiles corresponding to I1–I2, J1–J2, K1–K2, and L1–L2 labeled in b. Black, blue, and red lines correspond to times labeled in i. Dots show observed elevations, lines have tidal height subtracted. Due to >30 m d⁻¹ moving speed of ice in the study area, an Eulerian reference frame is not suitable for mélange thinning estimate, therefore a Lagrangian reference frame is used. m–o Inferred mean thinning rates from measurements separated by 6 days, color from dark blue to dark red corresponding to the first and last image pairs.
calving-like collapses within the mélangé (Fig. 8c1, e2). While the driving force for long-term mélangé motion did not change significantly, the coincidence of calving-like collapses in the mélangé and increases in mélangé speed near the elevation step-change likely reflects reduction in buttressing force at that downstream location, presumably caused by removal of thick mélangé ice. In contrast, mélangé >1.5 km upstream from the elevation step-change did not show significant speed perturbations following these calving-like collapse events (Fig. 8d1, d2). We surmise that the thick mélangé upstream from the elevation step-change is tightly packed, behaving essentially as an ice shelf (albeit a weak one), as suggested by previous work. In this way, calving-like collapses within the mélangé did not initially change the buttressing force at the glacier front significantly, perhaps reflecting rapid decay in stress transmission with distance.

Further lines of evidence support the hypothesis that the initial mélangé collapse events did not affect buttressing of the glacier front, but did affect nearby mélangé upstream from the elevation step-change. In strain rate maps (a more sensitive indicator of buttressing force change) calculated along the radar LOS (Methods), new extensional fissures appeared upstream from the elevation step-change immediately after each calving-like collapse, corresponding to the subsequent elevation step-change that would form during the next collapse event (Fig. 9 and Supplementary Fig. 8). If the buttressing force at newly formed elevation step-changes decreased significantly with each calving-like collapse event, the shear stress at the two sides of the fjord constraining the thick mélangé wedge must presumably have increased, in order to prevent rapid collapse of the remaining mélangé wedge. Further calving-like collapse events within the mélangé occurred between the end of our TRI observations and the first available Landsat-8 image (within 1.5 days), moving the elevation step-change closer to the glacier front (the fissure marked by the cyan arrow in Fig. 9d likely failed in the next calving-like collapse event after the TRI observation period). At some point the increased shear stress at the margins of the mélangé wedge exceeded the yield stress, leading to collapse of the remaining wedge, and removal (or significant reduction) of the buttressing force on the glacier front. At this point, major calving events can occur at the glacier terminus. Note that, within uncertainties, neither the LOS nor the horizontal glacier speeds and longitudinal strain rates near the calving front changed significantly either before or after the major calving events (Fig. 8b, Supplementary Figs. 7 and 8). This may reflect reconfiguration of the terminus as the glacier front retreats, and changes in flotation status. However, by the end of the TRI observation period, the glacier front close to the coast (dashed cyan outlined area of Fig. 9d) had more high strain rate zones than early in the TRI observations (Fig. 9a–c, Supplementary Figs. 7 and 8). This region of increased strain rate is also adjacent to the area that calved within 1.5 days after the end of the TRI observations (shaded blue in Fig. 9k). Combined with the progressive formation of newly formed fissures in the mélangé, this observation supports an overall decrease of mélangé buttressing force during the TRI observation period.

We have proposed that loss of the wedge-like mélangé immediately in front of the glacier contributes to renewed calving. We now attempt to quantify this effect in terms of changes in buttressing force, using two approaches (see Methods).

First, assuming that the mélangé acts as a weak granular material, buttressing from the downstream thin mélangé is a resistive force that prevents or limits calving-like collapses within the upstream mélangé wedge. This buttressing force will decrease to a low value (and possibly zero) immediately after each collapse and then increase until the next collapse. This is supported by our data: ice surface velocity immediately upstream from the elevation step-change increases stepwise after calving-like
collapses (Fig. 8e1); new extensional fissures formed in the mélange wedge immediately after each calving-like collapse (Fig. 9); DEMs show that ice thickness reaches a minimum immediately downstream from the elevation step-change (Fig. 10a). Under this condition, buttressing force from mélange downstream from the elevation step-change during periods of mélange quiescence approximates the buttressing force reduction acting on the remaining thick mélange wedge immediately after each calving-like collapse. This also represents reduction in buttressing force on the glacier front immediately before major calving on 20 or 21 June 2016, assuming the thick mélange wedge had retreated to a minimum by then. Applying the model of Burton et al.\textsuperscript{20} and simplifying the mélange as a cuboid (Methods and Supplementary Fig. 9), we estimate a buttressing force per unit lateral-width of $1.1 \times 10^7$ N m\textsuperscript{-1}. This is a minimum estimate of buttressing force decrease per meter of lateral-width during the TRI observation period is $-0.9$–$1.8 \times 10^7$ N m\textsuperscript{-1}, equal to $-11$–$22$ kPa pressure change on the entire glacier front assuming it has a thickness of 800 m. This buttressing force decrease will be even larger, reaching $-2.1$–$4.3 \times 10^7$ N m\textsuperscript{-1} (equal to a $-27$–$54$ kPa pressure change upon the entire glacier front assuming it has a thickness of 800 m) by the beginning of major calving events when further calving-like collapses moved the remaining mélange wedge away within 1.5 days after the end of TRI observation period (Methods).

Amundson and Burton\textsuperscript{33} modeled winter mélange at Jakobshavn Isbør as a quasi-static granular material and found a back-force from the mélange of order $1.0 \times 10^7$ N m\textsuperscript{-1} is sufficient to decelerate an already overturning iceberg or prevent an iceberg from overturning in the first place.\textsuperscript{20,33} A finite element model suggested that back-force of this magnitude is sufficient to reduce fracture propagation near

Fig. 7 Mélange ice loss and glacier calving rate. a–c Surface elevation maps at selected times. Dashed polygon P outlines a test area used to calculate ice loss, moving with the three vertices v1–v3 adjacent to the glacier (marked in (a)). Cyan rectangle in (a) marks an area whose elevations are shown in Fig. 10a. d Elevations of an Eulerian profile marked by cyan lines in (a–c). Large icebergs downstream from the white dot of (e) (shown with pink dots) were filtered for clarity. e Black dots show inferred ice mass within the test area; red line shows calving rate in daily increments.
the upper surface of the glacier front by reducing tensile stress here. Based on the >400 m maximum inferred thickness of the mélangé wedge during our observation period, when tightly packed mélangé extended down a significant fraction of the glacier front, we hypothesize that a thick mélangé wedge can also reduce growth and propagation of basal fractures (Fig. 10a, c). Elevation data at the same location and the same time of year in 2015 illustrate the contrasting scenario with a thin mélangé (Fig. 10b, d). Due to similarities in the speed and strain rate responses to calving-like events between the mélangé wedge and glacier, we cannot distinguish whether part of the glacier front was actually detached ice blocks whose rotations were inhibited by the presence of thick mélangé.

Our new observations yield direct support for the hypothesis that tightly packed mélangé can suppress iceberg calving. While this is consistent with previous research, our observations and analysis provide important new insights. In particular, these new data provide a quantitative framework to map tidel-timescale or shorter timescale elevation variations of pro-glacial mélangé and their influence on calving across the entire glacier front. To our knowledge, this is the first quantitative study of mélangé changes at daily and sub-daily timescale, and the first observation of a step-like boundary within the mélangé, separating low elevation, loosely packed downstream mélangé from a wedge of more tightly packed mélangé near the glacier front. Past estimates of mélangé thickness used in modeling either relied on limited data (characterised by low-spatial resolution and/or long revisit times) or assumed a uniform thickness for the mélangé. Our observations clearly show a distinct thickness change in the mélangé within a few kilometers of the glacier front during periods of suppressed calving. The TRI technique and our approach can be applied to other tidewater
**Fig. 9** LOS strain rate responses to calving-like collapses. Strain rates are calculated along line-of-sight directions, based on 30-min median LOS velocity maps derived from interferograms of adjacent radar measurements separated by 2 min. Cyan arrows mark newly formed fissures after calving-like collapse events. a, b are at 15 min before and after a major calving-like collapse on 15 June. Fissures in b evolved wider to c and failed on 20 June. The arrow in d marks a strain rate fissure by the end of TRI observation period. Dashed cyan in d outlines an area where new high strain rate zones appeared by the end of TRI observations, which is close to the calved area within 1.5 days after the TRI observation period (shaded in blue in Fig. 3k). Map area shown in this figure is outlined by the yellow box in Fig. 8a. Yellow lines mark the mélange-glacier boundary.

**Fig. 10** Mélange wedge and its impact on glacier calving. a, b Dark and light grey correspond to surface (measured) and bottom (inferred) heights in the outlined rectangle in Fig. 7a. distance is along profile from downstream. a is from this study (surface slope of the wedge is typically 0.2–2°) and b is from a 2015 campaign6,17 when the mélange is relatively thin. c, d Schematic configurations along flow section of the glacier terminus with and without the thick mélange. The thick, tapered mélange in a constrained channel buttresses the glacier front and reduces fracture propagation and calving.
grey dots in Supplementary Fig. 1c represent the heights (1-day median) of all points within a test area (white box 2 in Supplementary Fig. 1a) in the mélange, and show an obvious trend with distance. Assuming mélange ice is in gravitational-buoyancy equilibrium, no obvious trend should exist. This trend is possibly due to incomplete elimination of errors in the first-stage correction described above, because we simply used a linear correction model determined by limited near-field points. Two error sources can bias elevation estimate in the mélange: first, errors in elevations of the limited near-field points used in Supplementary Fig. 1b, second, incomplete removal of error due to imperfect vertical mounting of antennas and a linear term. The first type of error can be eliminated if more evenly distributed points with accurate elevations are available. For the second type of error, we note that our critical observation area is >2 km from the radar, where surface elevation varies at a level of 10 m of s meters, the nonlinearity will only cause errors at the cm level. However, the data used to fit the linear model in Eq. (1) are in the range 0.4–2.5 km from the radar, with up to ∼200 m height difference. A linear model based on these data can significantly bias the last term in Eq. (1), and then propagate through the entire DEM. We also note that the ArcticDEM is referenced to the WGS84 ellipsoid, leading to an offset between the height datum of TRI-derived DEMs and local sea level.

Both of the above two types of errors can be removed or minimized using measured heights within a relatively flat area in the mélange. We fit a plane to measured heights within box 2 (Supplementary Fig. 1a, downstream from the step-change in the mélange) that is closely parallel to local sea level (assuming mélange ice in the box maintains a flat surface defined by fixed water). We use an iterative least squares method to fit the plane whose distance to a point is the baseline error, and other sources of error in the derived DEMs, discussed below.

Two steps are needed to estimate a DEM from an unwarped phase map: first, to estimate the offset between unwarped phase at the elevation reference point and calculated phase based on Eq. (1); second, to estimate heights at points of interest based on the unwrapped phases from the first step. In the first step, elevation of the radar was measured with a single frequency GPS, and the resulting uncertainty can exceed 10 m. In addition, high-precision ground control points were not available. We, therefore, used radar and reference elevations estimated from the ArcticDEM data, provided by the Polar Geospatial Center at the University of Minnesota (https://www.pgc.umn.edu/data/arcticdem/). The absolute accuracy of ArcticDEM in this area has not been verified, however, this does not significantly affect the accuracy of our final estimate of TRI-derived DEMs, also discussed below.

For each unwrapped TRI phase map, when using the height difference of z (including uncertainty) between the radar and reference point to estimate the phase at the reference position, we could solve a quadratic equation of one unknown (6) based on Eq. (1). However, due to uncertainties in the ArcticDEM, baseline error, and imperfect vertical mounting of the interferometer, the phase estimate at the reference point may have an error. Also, baseline error and the tilt of the interferometer propagate into the Eq. (1) used to calculate the elevation maps. Ignoring terms that can cause error at very low levels (<1 cm), the uncertainty in the height estimate [13, 14]

\[
z = \frac{\lambda B}{2\pi R} \left[ 1 - \left( \frac{2}{\pi} \right)^2 \right] f_i
\]

where \(\lambda\) represents surface topography (height between radar and the study point), \(\lambda\) is the radar wavelength (1.74 cm), \(R\) is the range from the radar to the study point, \(B\) is the baseline length (distance between receiving antennas), and \(f_i\) is the phase. Depending on the application, a typical value for \(R\) is ∼25 km [13], representing a compromise between precision in the phase difference measurement (related to the DEM precision, where larger \(B\) values are preferred) and the ability to avoid phase breaks (phase unwrapping error, where smaller \(B\) values are preferred). In this study, we chose a relatively large \(B\) value (60 cm) and developed new approaches to minimize phase errors. The absolute accuracy of ArcticDEM in this area has not been verified, however, this does not significantly affect the accuracy of our final estimate of TRI-derived DEMs, also discussed below.

For each unwrapped TRI phase map, when using the height difference of \(z\) (including uncertainty) between the radar and reference point to estimate the phase at the reference position, we could solve a quadratic equation of one unknown (6) based on Eq. (1). However, due to uncertainties in the ArcticDEM, baseline error, and imperfect vertical mounting of the interferometer, the phase estimate at the reference point may have an error. Also, baseline error and the tilt of the interferometer propagate into the Eq. (1) used to calculate the elevation maps. Ignoring terms that can cause error at very low levels (<1 cm), the uncertainty in the height estimate [13, 14]

\[
\sigma_z = \frac{1}{2\pi R} [\sigma_{fi}^2 + \sigma_{fl}^2] = \frac{1}{2\pi R^2} \left[ R_0^2 + R_0^2 \right] f_i
\]

where \(\sigma_{fi}\) is phase error due to errors in the reference heights and instrumental geometry, \(\sigma_{fl}\) is noise in the radar measurement, \(\sigma_{fl}\) is baseline error, and \(\sigma_{fl}\) is the error caused by assuming the vertical axis of the three antennas is perfectly vertical, with an angle of \(\Phi\) from LOS direction. Note that except for random noise in the phase measurement, the other error sources are systematic, in the sense that they will cause similar errors that propagate across the entire elevation map. Ideally, with accurate knowledge of multiple ground control points, \(\sigma_{fl}\) and \(\sigma_{fl}\) and \(\sigma_{fl}\) in Eq. (2) can be explicitly solved and corrected [13]. However, high precision ground control points are not available in our study area. Typical error in the baseline determination is at the 0.1 cm level, and typical error in the tilt angle of antennas is at the 0.1° level [13]. These two error sources typically cause smaller errors compared to the phase error, therefore the dominant error is linearly dependent on \(R\). Thus we use a linear model to correct the majority of errors based on Eq. (2). Remaining error (e.g., the last term in Eq. (2) which is not linearly proportional to distance) will be discussed later.

Supplementary Fig. 1b shows the difference between the elevation estimates and the ArcticDEM for points near the radar (small white triangle and square, marked as 1 in Supplementary Fig. 1a). An obvious trend can be seen, indicating possible errors in the ArcticDEM, the baseline, or front-back tilt of the rack structure that supports the antennas. We excluded side-to-side tilt because it would cause a correction error on the different interferograms. The antenna rack was mounted on stable rock, the antennas were bubble-leveled, and the system was protected from wind by a radome. We thus use a simple 1-D model based on the best fitting line of \(RMS\) vs. slant range (red line in Supplementary Fig. 1b) to correct errors related to reference elevations and instrument geometry. Note that this assumes no systematic spatial bias in the ArcticDEM. Ideally, more evenly distributed points with known a priori elevations should be used as references for this correction. However in our case, only limited stationary areas were in the radar view.

Methods

DEM generation and uncertainty assessment. The TRI we use has one transmitting and two receiving antennas [13, 14]. To generate DEMs, data are collected by both receiving antennas to form interferograms. Assuming the interferometer is vertical, unwarped phases can be converted to elevations [13] using

\[
z = \frac{\lambda B}{2\pi R} \left( 1 - \left( \frac{2}{\pi} \right)^2 \right) f_i
\]

where \(\lambda\) represents surface topography (height between radar and the study point), \(\lambda\) is the radar wavelength (1.74 cm), \(R\) is the range from the radar to the study point, \(B\) is the baseline length (distance between receiving antennas), and \(f_i\) is the phase. Depending on the application, a typical value for \(R\) is ∼25 km [13], representing a compromise between precision in the phase difference measurement (related to the DEM precision, where larger \(B\) values are preferred) and the ability to avoid phase breaks (phase unwrapping error, where smaller \(B\) values are preferred). In this study, we chose a relatively large \(B\) value (60 cm) and developed new approaches to minimize phase errors. The absolute accuracy of ArcticDEM in this area has not been verified, however, this does not significantly affect the accuracy of our final estimate of TRI-derived DEMs, also discussed below.

For each unwrapped TRI phase map, when using the height difference of \(z\) (including uncertainty) between the radar and reference point to estimate the phase at the reference position, we could solve a quadratic equation of one unknown (6) based on Eq. (1). However, due to uncertainties in the ArcticDEM, baseline error, and imperfect vertical mounting of the interferometer, the phase estimate at the reference point may have an error. Also, baseline error and the tilt of the interferometer propagate into the Eq. (1) used to calculate the elevation maps. Ignoring terms that can cause error at very low levels (<1 cm), the uncertainty in the height estimate [13, 14]

\[
\sigma_z = \frac{1}{2\pi R} [\sigma_{fi}^2 + \sigma_{fl}^2] = \frac{1}{2\pi R^2} \left[ R_0^2 + R_0^2 \right] f_i
\]

where \(\sigma_{fi}\) is phase error due to errors in the reference heights and instrumental geometry, \(\sigma_{fl}\) is noise in the radar measurement, \(\sigma_{fl}\) is baseline error, and \(\sigma_{fl}\) is the error caused by assuming the vertical axis of the three antennas is perfectly vertical, with an angle of \(\Phi\) from LOS direction. Note that except for random noise in the phase measurement, the other error sources are systematic, in the sense that they will cause similar errors that propagate across the entire elevation map. Ideally, with accurate knowledge of multiple ground control points, \(\sigma_{fl}\) and \(\sigma_{fl}\) and \(\sigma_{fl}\) in Eq. (2) can be explicitly solved and corrected [13]. However, high precision ground control points are not available in our study area. Typical error in the baseline determination is at the 0.1 cm level, and typical error in the tilt angle of antennas is at the 0.1° level [13]. These two error sources typically cause smaller errors compared to the phase error, therefore the dominant error is linearly dependent on \(R\). Thus we use a linear model to correct the majority of errors based on Eq. (2). Remaining error (e.g., the last term in Eq. (2) which is not linearly proportional to distance) will be discussed later.

Supplementary Fig. 1b shows the difference between the elevation estimates and the ArcticDEM for points near the radar (small white triangle and square, marked as 1 in Supplementary Fig. 1a). An obvious trend can be seen, indicating possible errors in the ArcticDEM, the baseline, or front-back tilt of the rack structure that supports the antennas. We excluded side-to-side tilt because it would cause a correction error on the different interferograms. The antenna rack was mounted on stable rock, the antennas were bubble-leveled, and the system was protected from wind by a radome. We thus use a simple 1-D model based on the best fitting line of \(RMS\) vs. slant range (red line in Supplementary Fig. 1b) to correct errors related to reference elevations and instrument geometry. Note that this assumes no systematic spatial bias in the ArcticDEM. Ideally, more evenly distributed points with known a priori elevations should be used as references for this correction. However in our case, only limited stationary areas were in the radar view.
to fit the RMS vs. slant range scatter. The grey curve in Supplementary Fig. 1f is the best fitting curve to the dots with light color. Black is the best fitting curve to the scatter calculated from 30-min median and S2T median elevation time series. The fitting curves describe RMS of the majority of points quite well, although they may be biased at some locations, especially at the areas adjacent to TRI LOS shadow. For the black curve in Supplementary Fig. 1f, the coefficient $a = -0.02 \text{ km}^{-2}$ (for convenience, $R$ has units of km in Eq. (4), while RMS has units of m). We use this for error propagation in our melt rate and ice mass loss estimates (below) since these are all based on 30-min median DEMs. Supplementary Fig. 1g shows elevation time series for representative points marked by the same color dot within corresponding boxes in Supplementary Fig. 1a, light and dark colors represent non-smoothed and 30-min median filtered time series. Note that 30 min is a reasonable window because ice in the meltage moves at a speed of ~30–50 m d⁻¹, thus ice motion within a typical 30-min period is of order <0.1 pixel width (1 pixel width is 10 m in these TRI images), similar as Xie et al.6. Tidal measurements observed from a mooring at the mouth of Kangerdlugssuaq fjord in 2015 show no significant difference compared to tidal predictions is significant larger than the tidal model, and the phase difference is also larger than the extracted tides from elevation data shown in Supplementary Fig. 3c. We interpret the TRI observed displacement (grey dots in Supplementary Fig. 3f) onto vertical direction using vertical motion (Supplementary Fig. 3c describes the geometry), we can calculate vertical motion by inversely projecting the integrated LOS displacements (black dots in Supplementary Fig. 3d) onto vertical direction using

$$D_{\text{los}} = \frac{\rho_{\text{w}}}{\rho_{\text{w}} + \rho_{\text{fl}}} D_{\text{v}}$$

where $D_{\text{los}}$ is integrated vertical displacement (tidal variation) $R$ is slant range distance from the radar to the study point, ~2.8 km, $H_{\text{f}}$ is the height difference between the radar and the study point, ~190 m, $D_{\text{v}}$ is integrated LOS displacement measured by TRI. These parameters are shown in Supplementary Fig. 3c.

Red dots in Supplementary Fig. 3f show elevation time series (frequencies between 0.8 and 4 cycle per day passed) from Eq. (7). The overall amplitude of these tidal estimates is significantly larger than the tidal model, and the phase difference is also larger than the extracted tides from elevation data shown in Supplementary Fig. 3c. We interpret the TRI observed displacement (grey dots in Supplementary Fig. 3f) onto vertical direction using vertical motion (Supplementary Fig. 3c describes the geometry), we can calculate vertical motion by inversely projecting the integrated LOS displacements (black dots in Supplementary Fig. 3d) onto vertical direction using

$$D_{\text{los}} = \frac{\rho_{\text{w}}}{\rho_{\text{w}} + \rho_{\text{fl}}} D_{\text{v}}$$

where $D_{\text{los}}$ is integrated vertical displacement (tidal variation) $R$ is slant range distance from the radar to the study point, ~2.8 km, $H_{\text{f}}$ is the height difference between the radar and the study point, ~190 m, $D_{\text{v}}$ is integrated LOS displacement measured by TRI. These parameters are shown in Supplementary Fig. 3c.

Red dots in Supplementary Fig. 3f show elevation time series (frequencies between 0.8 and 4 cycle per day passed) from Eq. (7). The overall amplitude of these tidal estimates is significantly larger than the tidal model, and the phase difference is also larger than the extracted tides from elevation data shown in Supplementary Fig. 3c. We interpret the TRI observed displacement (grey dots in Supplementary Fig. 3f) onto vertical direction using vertical motion (Supplementary Fig. 3c describes the geometry), we can calculate vertical motion by inversely projecting the integrated LOS displacements (black dots in Supplementary Fig. 3d) onto vertical direction using

$$D_{\text{los}} = \frac{\rho_{\text{w}}}{\rho_{\text{w}} + \rho_{\text{fl}}} D_{\text{v}}$$

where $D_{\text{los}}$ is integrated vertical displacement (tidal variation) $R$ is slant range distance from the radar to the study point, ~2.8 km, $H_{\text{f}}$ is the height difference between the radar and the study point, ~190 m, $D_{\text{v}}$ is integrated LOS displacement measured by TRI. These parameters are shown in Supplementary Fig. 3c.

Red dots in Supplementary Fig. 3f show elevation time series (frequencies between 0.8 and 4 cycle per day passed) from Eq. (7). The overall amplitude of these tidal estimates is significantly larger than the tidal model, and the phase difference is also larger than the extracted tides from elevation data shown in Supplementary Fig. 3c. We interpret the TRI observed displacement (grey dots in Supplementary Fig. 3f) onto vertical direction using vertical motion (Supplementary Fig. 3c describes the geometry), we can calculate vertical motion by inversely projecting the integrated LOS displacements (black dots in Supplementary Fig. 3d) onto vertical direction using

$$D_{\text{los}} = \frac{\rho_{\text{w}}}{\rho_{\text{w}} + \rho_{\text{fl}}} D_{\text{v}}$$

where $D_{\text{los}}$ is integrated vertical displacement (tidal variation) $R$ is slant range distance from the radar to the study point, ~2.8 km, $H_{\text{f}}$ is the height difference between the radar and the study point, ~190 m, $D_{\text{v}}$ is integrated LOS displacement measured by TRI. These parameters are shown in Supplementary Fig. 3c.

Red dots in Supplementary Fig. 3f show elevation time series (frequencies between 0.8 and 4 cycle per day passed) from Eq. (7). The overall amplitude of these tidal estimates is significantly larger than the tidal model, and the phase difference is also larger than the extracted tides from elevation data shown in Supplementary Fig. 3c. We interpret the TRI observed displacement (grey dots in Supplementary Fig. 3f) onto vertical direction using vertical motion (Supplementary Fig. 3c describes the geometry), we can calculate vertical motion by inversely projecting the integrated LOS displacements (black dots in Supplementary Fig. 3d) onto vertical direction using

$$D_{\text{los}} = \frac{\rho_{\text{w}}}{\rho_{\text{w}} + \rho_{\text{fl}}} D_{\text{v}}$$

where $D_{\text{los}}$ is integrated vertical displacement (tidal variation) $R$ is slant range distance from the radar to the study point, ~2.8 km, $H_{\text{f}}$ is the height difference between the radar and the study point, ~190 m, $D_{\text{v}}$ is integrated LOS displacement measured by TRI. These parameters are shown in Supplementary Fig. 3c.
level defined by us (Supplementary Fig. 1c, d) may cause an offset of these values. Ice density can also differ due to variable compaction, while surface water density can differ due to changes in salinity, which may not be perfectly incompressible. However, these will not change the signs of Eqs. (8) and (9) or adversely change the mechanisms of overall ice loss in the mélange, and will not affect our major conclusions.

Glacier calving rate. Glacier calving events are determined by inspection of TRI, Landsat-8, and Sentinel-2/1 images. Figure 3 lists selected images to show ice loss due to glacier calving between 5 May and 29 June 2016. Before and after the period shown in Fig. 3c–i, there are additional calving events visible on satellite images. We chose this period to derive a relation between changes in the mélange and glacier calving because we have TRI observations within this period.

To estimate ice loss due to calving, we first digitize calving front positions by manually drawing locations of the glacial front on different images. 3-pixel width automatically drawn calving fronts are checked by digitizing and geolocation. From 21 May 2016 until the end of TRI observations, only one minor glacier calving event was detected (Fig. 3g, g). Large calved areas are found after the TRI observations (Fig. 3j–l).

The area of glacier ice loss has two components: first, changes between the digitized ice cliff positions, shaded with blue or red in Fig. 3k, l; second, ice that is out of the shaded areas before the calving events but later falls within the shaded areas due to ice motion (ideally, calved ice area should be estimated in a Lagrangian reference frame). The first part accounts for the majority of total glacier ice loss during our study period. The second part accounts for ~10% because ice near the glacier front moves fast (~30 m d−1). Ice loss due to this component is calculated using velocities from feature tracking and is shown in Supplementary Fig. 6.

To convert area of ice loss into mass of ice loss, we assume an average ice thickness of 800 m. This is based on measured surface elevation (~100 m) (Fig. 2) and the depth of bed benthymetry (~600–1200 m) near the glacier front. For the period 20–29 June, total calved ice mass is 6.7 ± 0.8 Gt. For reference, average annual mass loss for all of Greenland, measured for the decade 2003–2013 was 244 ± 6 Gt (20). A different ice thickness will change the values of calved ice mass and calving rate (we define it as calved ice mass per day), but would not change the inverse relation between mélange ice mass and glacier calving rate (Fig. 7a).

The mass of the wedge of mélange ice in front of the glacier grows by calving, and shrink by downstream advection and divergence of ice, gravitational collapse of elevated ice at the toe of the wedge, and sub-aerial and submarine melting due to local changes in surface elevation. While the ice over the mélange wedge may be offset by local changes in surface elevation, the layer of ice on the mélange wedge will move in response to changes in surface elevation. If the surface elevation change is only locally significant, then the glacial ice may slide, which is addressed in the next section. If the surface elevation change is significant, then thinning of the mélange wedge may occur, and the mélange wedge may move in response to changes in surface elevation.

Speed and strain rate changes. Supplementary Fig. 7a, c, e shows examples of speed changes before and after major calving events, estimated by feature tracking. Supplementary Fig. 7b, d, i shows longitudinal strain rates during different periods calculated by using the logarithmic strain-rate calculation code of Alley et al. (2014) with an effective length scale of 300 m. Different length scales do not change the overall pattern but longer length scales yield smoother strain rate maps. Despite changes in mélange thickness, terminus position, and possibly fluctuation status, speed and longitudinal strain rate in the middle of the glacier show no significant increase. By the end of the TRI observation period, the glacier front near the coast (to the radar side) has higher speed and longitudinal strain rate compared to the beginning of the TRI observation period, corresponding to the area with newly formed fissures. The area of mélange between the glacier terminus and the outer mélange (Supplementary Fig. 8), suggesting a decrease in buttressing force from mélange downstream from the elevation step-change.

The longitudinal and transverse strain rates also provide a way to estimate divergence thinning rate. For the dashed cyan area in Supplementary Fig. 7b, d, i, we calculate divergence thinning rate of 0.044 d−1 on the first TRI observation day, and a divergence thinning rate of 1.84 m d−1 on the last TRI observation day. These are comparable to divergence rates estimated in Fig. 5a, and indicate an overall increase in ice mobility of the mélange. Supplementary Figure 8 shows LOS strain rate changes throughout the TRI observation period. For two points (P1 and P2) separated by a distance of d along one radar LOS, in a time interval of Δt, unwrapped phases at P1 and P2 changed by Δϕ1 and Δϕ2, respectively, then LOS strain rate between P1 and P2 during Δt is calculated as

$$\varepsilon_{los} = \frac{\lambda (\Delta \phi_2 - \Delta \phi_1)}{2dM} \quad (10)$$

where λ is the radar microwave wavelength (1.74 cm). Note the constant high and low zones (dark blue) can be caused by high gradients in LOS velocities due to geometric features. However, changes in LOS strain rate should represent real changes in strain rates. During the observation period, newly formed high LOS strain rate zones (marked by cyan arrows) occurred immediately after calving-like collapses, located upstream from the elevation step-changes.

Mélange buttressing force estimate. Two approaches have been used to estimate mélange buttressing force decrease, described below.

In the first method, we consider the mélange wedge as a weak ice shelf (20) that is an extension of the glacier, then buttressing from the downstream thin mélange is a resistive force that prevents calving. This is similar to the effect of the upstream mélange wedge, and also acts on the glacier front. The change of buttressing force at the elevation step-change is thus a lower bound estimate of change at the glacier front. We assume that the downstream mélange will decrease to a low value (and possibly zero) immediately after each collapse and then increases until the next collapse. This is supported by our data: ice speed immediately upstream from the elevation step-change increases stepwise after calving-like collapses (Fig. 8c), new longitudinal fissures formed in the mélange wedge immediately after each calving-like collapse (Fig. 9), DEMs here show that ice thickness immediately downstream from the elevation step-change is a minimum (Fig. 10a and Supplementary Fig. 4f). With this assumption, the change in buttressing force due to the change in mélange buttressing force (from 2019 to 2021) is estimated by feature tracking and is shown in Supplementary Fig. 6. Based on the thickness measurements of the mélange, we simulate the mélange as a cuboid 7.7 km wide, 31.0 km long, and 39.8 m thick (Supplementary Fig. 9). The thickness is an average of all pixels within the dashed polygon M in Supplementary Fig. 9b. Due to local changes in surface elevation, we simulate the mélange as a cuboid 7.7 km wide, 31.0 km long, and 39.8 m thick. The change of buttressing force due to the elevation step-change is estimated as a function of the elevation step-change (Fig. 8e1); new extensional strain rate zones (marked by cyan arrows) occurred immediately after calving-like collapses, located upstream from the elevation step-changes.
width applied on the calving face by the melt ice

\[ F_m = \sigma \Delta H_p \]  

(13)

During the ~13 day TRI observation period, average mélange thickness of the test area decreased by ~37 m. Substituting \( \Delta H_p = 37 \) m into (13) with the thickness decrease yields a total force decrease applied on the glacier front of \( -0.9 \times 10^7 \text{ N m}^{-1} \) (force per meter of lateral-width), equal to a ~11-22 kPa pressure change upon the entire glacier front assuming it has a thickness of 800 m. Further calving-like collapse events within the mélange between 12:20 20 June 2016 (fast TRI observation) and 00:00 22 June 2016 (first Landsat-8 image acquisition following the end of the TRI observations) may have reduced the buttressing force even more. If the entire mélange wedge was moved away by further calving-like collapses before major calving events, the total decrease of mélange thickness in the test area is ~89 m, yielding a total decrease of buttressing force by ~2.1-4.3 \times 10^7 \text{ N m}^{-1} before major calving events after the TRI observation period, equal to a ~27–54 kPa pressure change upon the entire glacier front assuming it has a thickness of 800 m. Since glacier speed did not change significantly during the TRI observation period, it is possible that most of the butte~ing force decrease acting directly at the glacier front occurred after removal of the remaining mélange wedge (within 1.5 days after the TRI observations). Perhaps buttressing force acting on the glacier remained largely unchanged during the TRI observations, reflecting rapid decay of stress transmission with distance, but buttressing force acting on the newly formed elevation step-change dropped significantly and shear stress at the constraining margin of the remaining mélange wedge increased. Once it reached a threshold (yield stress), the remaining mélange wedge failed, the glacier front became the new elevation step-change, buttressing force on the glacier calving face dropped to very low values, and large calving events occurred. Note that the average mélange thickness of our test area (Polygon P shown in Fig. 7a) may not be the best estimate of effective buttressing thickness, or properly account for the effects of wedge geometry. The threshold is ~0.8–4 \times 10^7 \text{ N m}^{-1} is likely a minimum estimate of the decrease in buttressing force from the beginning of TRI observations to the first major calving event. Further research is needed to quantify these effects.

Data availability

Landsat images were downloaded through the USGS EarthExplorer (https://earthexplorer.usgs.gov). Sentinel-2 data provided by European Space Agency and were downloaded through the USGS EarthExplorer and Alaska Satellite Facility (https://www.asf.alaska.edu/). TRI data (~2TB) are available upon reasonable request. Additional data can be found in the supplementary figures and movies. Programs and methods presented in this paper are either noted using references to their sources or provided with detailed equations.

Code availability

The codes used for data analysis and figure plotting are available from the corresponding author upon reasonable request.

Received: 14 June 2018 Accepted: 5 June 2019

Published online: 19 July 2019

References


Acknowledgements
Denise Holland at the Center for Global Sea Level Change in New York University Abu Dhabi organized the field logistics. Fanghui Deng at the University of South Florida is thanked for helpful discussions. This research was partially supported by NASA grant NNX12AK29G to T.H.D. D.M.H. acknowledges support from NYU Abu Dhabi grant G1204, NSF award ARC-1304137, and NASA Oceans Melting Greenland NNX15AD55G. ArcticDEM provided by the Polar Geospatial Center under NSF-OPP awards 1043681, 1509691, and 1542736.

Author contributions
S.X. analyzed the data and wrote the paper together with T.H.D. D.M.H. organized the TRI campaign and started a discussion that led to this research. D.V. operated the radar and helped in the initial data analysis. I.V. participated in discussions. All authors commented on the paper.

Additional information
Supplementary Information accompanies this paper at https://doi.org/10.1038/s41467-019-10908-4.

Competing interests: The authors declare no competing interests.

Reprints and permission information is available online at http://npg.nature.com/reprintsandpermissions/

Peer review information: Nature Communications thanks Martin Sharp and other anonymous reviewer(s) for their contribution to the peer review of this work. Peer reviewer reports are available.

Publisher’s note: Springer Nature remains neutral with regard to jurisdictional claims in published maps and institutional affiliations.
Appendix IV: Licence and reprint of Xie et al., 2019, IGR

3/16/2020

University of South Florida Mail - Request for a license for a published paper in IGR

Surui Xie <suruixie@mail.usf.edu>

Request for a license for a published paper in IGR

Academic UK Non Rightslink <permissionrequest@tandf.co.uk>
To: "suruixie@mail.usf.edu" <suruixie@mail.usf.edu>
Cc: "Yang, Maggie" <Maggie.Yang@tandfchina.com>

Mon, Mar 16, 2020 at 7:16 AM

Dear Surui Xie


Thank you for your correspondence requesting permission to reproduce your authors accepted manuscript from our Journal in your printed thesis and to be posted in the university’s repository – University of South Florida.

We will be pleased to grant permission on the sole condition that you acknowledge the original source of publication and insert a reference to the article on the Journals website: http://www.tandfonline.com

This is the authors accepted manuscript of an article published as the version of record in International Geology Review © 2019 Informa UK Limited, trading as Taylor & Francis Group, and insert a link to the articles Version of Record

This permission does not cover any third party copyrighted work which may appear in the material requested.

Please note that this license does not allow you to post our content on any third party websites or repositories.

This licence does not allow the use of the Publishers version/PDF (this is the version of record that is published on the publisher’s website) to be posted online.

Thank you for your interest in our Journal.

Yours sincerely

Karin Beesley - Permissions Administrator, Journals
Taylor & Francis Group
3 Park Square, Milton Park, Abingdon, Oxon, OX14 4RN, UK

Permissions Tel: +44 (0)20 7017 7617
Permissions e-mail: permissionrequest@tandf.co.uk

Taylor & Francis Group is a trading name of Informa UK Limited, registered in England under no. 1072954
A new geological slip rate estimate for the Calico Fault, eastern California: Implications for geodetic versus geologic rate estimates in the Eastern California Shear Zone

Surui Xie*, Elisabeth Gallant#, Paul H. Wetmore#, Paula M. Figueiredo#, Lewis A. Owen#, Craig Rasmussen*, Rocco Malservisi*, Timothy H. Dixon#

*School of Geosciences, University of South Florida, Tampa, FL, USA; #Department of Geology, University of Cincinnati, Cincinnati, OH, USA; †Department of Soil, Water and Environmental Science, University of Arizona, Tucson, AZ, USA

CONTACT Surui Xie: suruixie@mail.usf.edu School of Geosciences, University of South Florida, Tampa, FL, USA

Accurate estimation of fault slip rate is fundamental to seismic hazard assessment. Previous work suggested a discrepancy between short-term geodetic and long-term geologic slip rates in the Mojave Desert section of the Eastern California Shear Zone (ECSZ). Understanding the origin of this discrepancy can improve understanding of earthquake hazard and fault evolution. We measured offsets in alluvial fans along the Calico Fault near Newberry Springs, California, and used several techniques to date the offset landforms and determine a slip rate. Our preferred slip rate estimate is 3.2±0.4 mm/yr, representing an average over the last few hundred thousand years, faster than previous estimates. Seismic hazard associated with this fault may therefore be higher than previously assumed. We discuss possible biases in the various slip rate estimates and discuss possible reasons for the rate discrepancy. We suggest that the ECSZ discrepancy is an artifact of limited data, and represents a combination of faster slip on the Calico Fault, off-fault deformation, unmapped fault strands, and uncertainties in the geologic rates that have been underestimated. Assuming our new rate estimate is correct and a modest amount (40%) of off-fault deformation occurs on major ECSZ faults, the summed geologic rate estimate across the Mojave section of the ECSZ is 10.5±3.1 mm/yr, which is equivalent within uncertainties to the geodetic rate estimate.

Keywords: Calico Fault; geological slip rate; geodetic slip rate; Eastern California Shear Zone; displacement; surface exposure dating

1. Introduction

The slip rate of an active fault is a fundamental parameter in seismic hazard estimation (Petersen et al., 2015). Knowledge of strain partitioning and slip rate accommodation across a plate boundary zone is also important for understanding how faults evolve and interact with other faults (Dolan et al., 2007; Ye and Liu, 2017; Dixon and Xie, 2018). Fault slip rate can vary in both space and time, potentially affecting the timing and magnitude of future damaging earthquakes, emphasizing the importance of detailed studies.

The Eastern California Shear Zone (ECSZ) accommodates ~20–25% of Pacific-North America plate motion in central and southern California, northeast of the Big Bend of the San Andreas Fault (Dokka and Travis, 1990a, b; Sauber et al., 1994; Dixon et al., 1995, 2000; Miller et al., 2001; Lifton et al., 2013; Figure 1). Most of the remaining plate motion is accommodated to the west, on the San Andreas Fault in central California, or the San Andreas, San Jacinto and Elsinore faults in southern California (e.g., Bennett et al., 1996; Meade and Hager, 2005; Shen et al., 2011). Formation of the ECSZ is kinematically linked to the Big Bend, whose formation in turn is related to the inland jump of the southern part of the plate boundary at ~5–10 Ma (Atwater and Stock, 1998; McQuarrie and Wernicke, 2005). Several faults within the ECSZ likely formed or accelerated around this time or later (Dokka and Travis, 1990a, b).

The region has been an important natural laboratory to study the formation and evolution of faults (Frankel et al., 2008), as well as other tectonic and plate kinematic studies. Dokka and Travis (1990a, b) recognized the importance of accommodating a significant fraction of Pacific-North America plate motion. Minster and Jordan (1987) first identified the ‘San Andreas discrepancy’; the discrepancy represents the difference between overall plate motion and motion carried by the San Andreas Fault. The discrepancy was initially attributed to significant right-lateral shear on other faults within the Basin and Range province to the east and the California continental margin to the west (Minster and Jordan, 1987; Ward, 1990). Improvements in geodetic data in the last few decades have clarified slip partitioning across the entire Pacific-North...
America plate boundary, and suggest a general agreement between summed geodetic slip rates across individual deforming zones and overall relative plate motion (e.g., DeMets and Dixon, 1999; Sella et al., 2002; Kreeemer et al., 2003; DeMets and Merkouriev, 2016).

More recently, several researchers (e.g., Gan et al., 2000; Meade and Hager, 2005; Oskin et al., 2008; Spinler et al., 2010; Evans et al., 2016) have noted discrepancies between geologically determined and geodetically determined slip rate estimates for individual faults within the ECSZ, or for summed rates across shear zone, hereafter termed the ECSZ discrepancy. For example, in the Mojave Desert region (Figure 1) the summed geologic slip rate across the region at ~34.8°N has been defined as ≤6.2±1.9 mm/yr (Oskin et al., 2008), while geodetic rate estimates are significantly faster, ~11 to ~18 mm/yr (Evans et al., 2016 and references therein) (Figure 2). A variety of factors could contribute to the ECSZ discrepancy, including:

1. Off-fault deformation, such that fault slip rates *sensu stricto* are less than the integrated block motion rate across the larger fault zone (e.g., Shelef and Oskin, 2010; Dolan and Haravitch, 2014; Herbert et al., 2014a).

2. Acceleration of young, immature faults (Gourmelen et al., 2011), such that the geologic rate, which may average over the early stages of a fault zone’s activity, will be less than the current rate measured by geodesy.

3. Temporal changes in fault slip rates beyond simple acceleration, reflecting complex tectonic processes in the ECSZ, including transient strain on individual faults or temporally clustered earthquakes at the scale of the shear zone (Rockwell et al., 2000; Peltzer et al., 2001; Oskin and Iriondo, 2004; Meade and Hager, 2005; Dolan et al., 2007; Oskin et al., 2007a, 2008; Cooke and Dair, 2011; Dixon and Xie, 2018).

4. The effects of post-seismic motion and visco-elastic relaxation, such that geodetic rates within a few decades of a major earthquake are faster than their long-term average (Dixon et al., 2003; Chuang and Johnson, 2011; McGill et al., 2015). In other words, the rates differ only because the long-term rate is not properly modeled in some geodetic approaches, for example, those that assume purely elastic rheology.

5. Unmapped faults.

6. Systematic errors in one or both of the geodetic and geologic techniques.

Determining the origin of such discrepancies, in the ECSZ and elsewhere, is important for a variety of reasons, including improved understanding of earthquake process and fault evolution, as well as seismic hazard assessment. In its simplest form, the potential seismic hazard for a given fault is positively correlated with the fault’s slip rate: in a given period of time, faults with higher slip rates are loaded faster than faults with lower slip rates (Petersen et al., 2015). Accurate estimation of a fault’s current slip rate, and possible long-term variation, is paramount.

Geological slip rate estimates suffer from a limited database. Bird (2007) investigated slip rate data for >800 faults in the conterminous western United States and found that only a small portion (~6%) have well-constrained rates (having combined probability density functions for long-term slip rate in which the width of the 95% confidence range is smaller than the median). He argued that ~4 offset features are required to achieve a well-constrained rate, and ≥7 offset features are required to guarantee a high degree of certainty. To our knowledge no fault in the Mojave ECSZ region has been studied sufficiently to meet Bird’s (2007) criteria and generate the necessary ensemble of rate estimates (Oskin et al., 2008 and references therein).

There are two other issues relevant to slip rate characterization: 1) Surface displacement can be heterogeneous along a fault, perhaps representing interactions between neighboring faults or different levels of off-fault deformation (e.g., Fletcher et al., 2014; Dolan and Haravitch, 2014); and 2) precise dating of offset features can be challenging, especially for Pleistocene and younger alluvial fans (a common offset marker), where surface exposure dating techniques typically exhibit a high degree of scatter. Consequently, a given slip-rate estimate may not be robust, emphasizing the importance of additional studies.

Here, we review geodetic and geologic slip rate estimates for the region and report a new geological slip rate estimate for the Calico Fault, a major fault within the Mojave Desert section of the ECSZ. The new rate is significantly faster than previously determined geologic slip rates, and hence bears on the issue (and perhaps the reality) of the ECSZ discrepancy.
2. Previous work

2.1 Prior geodetic studies

Fault slip rate estimates based on geodetic data are model-dependent. Most models assume either a purely elastic rheology or an elastic layer overlying on one or more visco-elastic layers. The latter has been used to study earthquake-cycle effects in several parts of the Pacific-North America boundary zone (Malservisi et al., 2001, 2003; Schmalzle et al., 2005; Fulton et al., 2010; Chuang and Johnson, 2011). Dixon et al. (2003), McGill et al. (2015) and Evans et al. (2016) suggested that discrepancies between geodetic and geologic slip rates in the ECSZ and Walker Lane (the northern continuation of the ECSZ) could be caused by prior earthquakes which stimulate visco-elastic deformation in the lower crust and upper mantle that varies over the time scale of an earthquake cycle. Liu et al. (2015) used historical triangulation/trilateration observations before the 1992 Landers earthquake and GPS measurements after the Landers earthquake to recover the secular deformation field and differentiate post-seismic transients. They found that the 1992 Landers and 1999 Hector Mine earthquakes adversely affect GPS measurements, with 2–3 mm/yr excess right-lateral shear inferred across the co-seismic ruptures in the post-earthquake GPS solutions. They estimate a cumulative long-term deformation rate of 13.2–14.4 mm/yr across the Mojave section of the ECSZ, similar within uncertainties to the pre-Landers geodetic estimate of 12 mm/yr by Sauber et al. (1994).

Herbert et al. (2014) used a boundary element method to simulate three-dimensional deformation of the ECSZ. Their modeling approach suggests that a block-like fault network (faults are simplified to be connected) can produce a cumulative strike-slip rate estimate that is 36% greater than a discontinuous fault model. Based on gradients in the derived deformation map and the implied strain energy density, Herbert et al. (2014a) concluded that 40±23% of the total strain across the ECSZ could be attributed to off-fault deformation.

Evans et al. (2016) used a total variation regularization method to investigate the role of fault system geometry in block models, determining a best-fitting geometry from an initial model with numerous faults. This method minimizes the influence of fault geometry assumptions and reduces uncertainties in geodetic slip rate estimates. Moreover, since a dense fault geometry was used in the initial model, which included active faults separated by <10 km, this modeling method should be able to assess the role of distributed deformation. Evans et al. (2016) identified persistent discrepancies between geologically and geodetically estimated slip rates in the ECSZ, with 4–7 mm/yr discrepancies on the Calico Fault. This suggests the importance of additional studies of the Calico Fault.

To the north of the ECSZ in the southern Walker Lane, across the Northern Death Valley-Fish Lake Valley Fault (DV-FLVF) and the White Mountain Fault (WMF), Lifton et al. (2013) compared GPS-based crustal velocities and geologic slip rates, and found that most of the observed discrepancy between long- and short-term slip rates occurs across Owens Valley. They concluded that the observed geodetic versus geologic discrepancy across the southern Walker Lane is likely a combination of under-estimated geologic slip rates on the WMF and broadly distributed deformation in Owens Valley that is not well preserved in the geologic record.

2.2 Prior geologic studies

From analysis of paleoseismologic trench data and offset landforms along the Calico Fault near Newberry Springs, California, Ganef et al. (2010) found that strain release on the Calico Fault has been highly episodic over the past ~9,000 yr, reinforcing the suggestion that earthquakes in the ECSZ are clustered (e.g., Rockwell et al., 2000; Dolan et al., 2007). The geomorphic displacements in the paleoseismologic evidence along the Calico Fault imply that more than one large earthquake (Mw ≥ 7.0) can occur in each clustering time period (Ganef et al., 2010).

Based on a fault initiation time between ~10.6 and 5.5 Ma and a total of 65 km right-lateral displacement at 35° N (from offset Early Miocene markers and a reconstruction model), Dokka and Travis (1990a, b) estimated an integrated long-term slip rate for the ECSZ at 6–12 mm/yr. Assuming that the ECSZ is kinematically linked to the Big Bend in the San Andreas Fault and the inland jump of the plate boundary to the present Gulf of California, age constraints on the timing of the inland jump and the timing of the initiation of marine sedimentation in the northern and southern Gulf of California allow refinement of this estimate. Oskin and Stock (2003) dated marine incursion into the southern Gulf of California at 8.2 Ma, while the northern Gulf is somewhat younger, 6.3–6.5 Ma, perhaps constraining the rate of northward propagation of the developing rift. Bennett et al. (2015) refined the age estimate of marine incursion into the northern Gulf, dating it at 6.2±0.2 Ma. The ECSZ likely formed or accelerated shortly after this time. Initiation ages for the ECSZ between 5.0 and 6.0 Ma allow for the possibility of some finite period for northward propagation. Using the 65 km displacement estimate of Dokka and
Travis (1990a, b) and this range of initiation ages the long-term average rate for the ECSZ of 10.8–13.0 mm/yr, essentially identical to most of the geodetic estimates within uncertainties.

Oskin et al. (2007a, 2008) measured surface displacements across several alluvial fans and a lava flow with different ages, determining the slip rates of six dextral faults (Helendale, Lenwood, Camp Rock, Calico, Psgah, and Ludlow) across the ECSZ, with an overall rate of ≤6.2±1.9 mm/yr at ~34.8°N. The Calico Fault had the fastest slip rate in these studies, 1.8±0.3 mm/yr. This slip rate is based on a 56.4±7.7 ka old surface offset near Sheep Spring Wash in the northern Rodman Mountains. In a study area located ~8 km southeast of that of Oskin et al. (2007a), Selander (2015) estimated a 1.4 ±0.8 mm/yr slipp rate for the Calico Fault based on a 17.1 ±6±7.3 ka old surface offset southwest of the Rodman Mountains. Further north, Oskin and Iriondo (2004) estimated a slip rate of 0.5 mm/yr for the Blackwater Fault, north of the Calico-Blackwater Fault system. Selander (2015) interpreted such rate fluctuations as evidence for strain transfer from the Calico Fault onto other several nearby faults, with overall dextral slip of the ECSZ apparently decreasing to the northwest, to ≤2.6±1.9 mm/yr north of 35°N.

A highly disconnected fault network in the ECSZ could imply significant off-fault deformation (Herbert et al., 2014a, b; Selander, 2015), or some amount of slip on fault strands that have not been studied. Using the deflection of continuous planar markers and the rotation of paleomagnetic sites, Shelef and Oskin (2010) found that distributed deformation over zones of 1–2 km width accommodates 0 to ~25% of the total displacement, with most displacement occurring within 100–200 m of faults, decreasing nonlinearly away from the fault.

2.3 The role of fault maturity

Wesnousky (1988) found that the structural complexity of strike-slip faults decreased with increasing offset. Several studies of ECSZ faults (e.g., Stirling et al., 1996; Rockwell et al., 2000; Dolan and Haravitch, 2014; Selander, 2015) emphasize their structural complexity, a key marker of immaturity, and one that could affect the measurement and interpretation of geologically-defined slip rates.

Wesnousky (2005) compared the San Andreas and ECSZ faults systems, noting the latter’s smaller cumulative offset and defining it as an immature fault system. Dolan and Haravitch (2014) analyzed a global set of large strike-slip earthquakes and found that faults with total offsets <25 km manifest only ~50–60% of earthquake slip as surface faulting, while faults with total offsets ≥ 85 km display ~85–95% of their slip on surface faults. Based on the inefficiency of generating surface faulting, the authors define strike-slip faults with ≤25 km total offset as immature faults. By this definition, all active faults in the Mojave section of the ECSZ are immature (Dokka, 1983; Dokka and Travis, 1990a; Dixon and Xie, 2018). For example, using well-defined markers from an early Miocene structural belt, Dokka (1983) estimated total offsets on individual active faults across the central Mojave Desert from 1.5 to 14.4 km. On the eastern margin of the ECSZ, the Bristol-Granite Mountain Fault has a total offset of 24 km (Lease et al., 2009), but it is currently inactive. The Calico Fault has a total offset of 9.8 km (Glazner et al., 2000; Oskin et al., 2007a).

While individual faults tend to lengthen, narrow and simplify with cumulative offset (Wesnousky, 1988; Dolan and Haravitch, 2014), plate boundary zones as a whole can grow in width and complexity (e.g., by adding new faults) depending on overall plate motion and other kinematic boundary conditions. Oldow et al (2008) describe evolution of the southern Walker Lane, generally considered the northern extension of the ECSZ, and note that parts of it have widened since the Pliocene.

3. New displacement observations

Our study area is located near Newberry Springs, California (Figure 1). Two alluvial fan surfaces here are offset by the Calico Fault (Figure 3). We refer to them as the Autumn Leaf Road (ALR) and Troy Road (TR) alluvial fans based on nearby roads. We estimated their strike-slip displacements based on three data sets: 1) our own field observations including mapped fault scarps; 2) high resolution aerial ortho-imagery with 0.3 m horizontal resolution, downloaded from USGS EarthExplorer: https://earthexplorer.usgs.gov/; and 3) a digital elevation model (DEM) derived from airborne LiDAR data with 0.5 m horizontal resolution and centimeter-level vertical precision, downloaded from OpenTopography Facility (http://www.opentopography.org/). Aerial ortho-imagery and LiDAR are especially useful in this semi-arid environment.
3.1 Autumn Leaf Road (ALR) alluvial fan

The ALR alluvial fan is the oldest observed alluvial fan surface in the Newberry Springs area and consists of a series of isolated alluvial surfaces elevated 2–6 m above younger surfaces (Figures 3–5). The surface of this alluvial fan is defined by a well-developed desert pavement, dominated by dark varnished pebbles and abundant, widely spaced meter-scale, sub-angular boulders with compositions that include quartzite, basalt, granite and rhyolite (Supplementary Figure S1). Fault traces identified by field mapping and aerial photography show a well-defined linear trace striking ~323°. The Calico Fault displaces the alluvial fan in a right-lateral sense, with the main body (better preserved) on the northeast wall and three smaller bodies (remnants, partially eroded) on the southwest wall, ~1 km to the northwest of the main body. The alluvial fan surface is characterized by shallow (0–2 m deep) channels that are partially filled by deposit and boulders, with one prominent (1–2 m deep) drainage on both the main body and the northwestern-most of the three smaller bodies (marked by red dots in Figure 4b). Excavation of two 2-m-deep, 1-m-wide, and 2-m-long trenches (CalicoA and Calico-Pit3, locations shown in Figure 3b) into the side of the ALR alluvial fan reveals that the deposit is dominated by cobbles and occasional boulders, with well-developed calcium carbonate coatings at depths >0.2 m. The coatings are ~4 mm (typically 0.5-2 mm) in thickness (Supplementary Figure S2), with some weak conjoining of adjacent clasts.

We reconstructed the pre-displacement ALR alluvial fan along the fault trace based on surface features and the LiDAR DEM, obtaining 1110±110 m (2σ uncertainty is used in this paper) of right-lateral displacement (Figures 4, 5). This places the largest of the three smaller alluvial fan bodies immediately adjacent to the southwestern margin of the main body. This also aligns the wide paleo-channel (between the cyan and yellow dots in Figure 4b) on both sides of the fault, and the prominent drainage on the alluvial fan surfaces (marked by red dots in Figure 4b). We note that this reconstruction aligns a paleo-creek strongly incised in the remnants of the alluvial fan in the northwest to the prominent drainage present on the southeastern alluvial fan body. This southeastern drainage starts deeply incised into the alluvial fan body but drains to the northeast, there is no topographic or geomorphologic reason for it, while the paleo-creek on the northwest alluvial fan body does not seem to have a continuation across the fault trace (Figures 3, 4). Thus, we interpret the prominent drainage as a pre-existing drainage that was once related to the dominant drainage on the northwester fan body. The width of the channel to the southeast of the major body (110 m) is used to define the uncertainty for this displacement, following the method of Frankel et al. (2011). We used several methods and tools to analyze the geomorphology in detail and interpret the displaced features, including the software package LaDiCaoz (v2.1) (Zielke et al., 2015; Haddon et al., 2016) (Figure 6). By shifting an elevation profile 25 m southwest from the fault (red line between two yellow dots in Figure 6a) along the fault trace, an 1111 m horizontal displacement minimizes misfit (Figure 6d) to the elevation profile northeast of the fault (solid blue line in Figure 6a), with a 27.5 m vertical displacement that is perhaps due to southwest-side down dipping of the fault (Selander, 2015). Figure 6f shows a restored contour map, where sharp-pointed V’s near the ALR alluvial fan are well-aligned between the two sides of the fault. Figure 6g and 6h show two elevation profiles (corresponding to red and blue dotted yellow lines in Figure 6a) along the dominant drainage before and after restoration; Figure 6i and 6j show two elevation profiles (corresponding to dashed red and blue lines in Figure 6a) on the fan surfaces before and after restoration. These show that the dominant drainage and alluvial fan surface on two sides of the fault have a similar slope aspect. Since the software package LaDiCaoz yields a displacement estimate that is virtually identical to the estimate based on reconstructions using aerial ortho-imagery and the LiDAR DEM, we use 1110 ±110 m as the displacement and corresponding uncertainty for the ALR alluvial fan. While using the selected paleo-channels as piercing points may induce uncertainty to the reconstruction of older alluvial fans due to erosion, we feel this is reflected in our uncertainty estimate.

3.2 Troy Road (TR) alluvial fan

The TR alluvial fan is located ~1 km southeast of the main body of the ALR alluvial fan (Figure 3). Here, rock varnish is moderately developed (light brown) on the surface, and desert pavement is not well developed, suggesting an age that is younger than the ALR alluvial fan, but older than the active channel. A network of partly filled channels and trains of 0.5–1-m-size boulders characterize the alluvial fan surface. Well-preserved bars with imbricated boulders are pervasive across this fan surface. Carbonate coatings and rubification of the undersides of clasts and boulders are indistinct or not developed within this fan on the surface and an evacuated trench (Calico-Pit2, location shown in Figure 3b), consistent with a younger age compared to the ALR alluvial fan. The TR alluvial fan is eroded along both its northwestern and southeastern margins, with active channels traversing the alluvial fan from south-southwest to northeast (Figures 7a, 8a). Surface fault traces show that the Calico Fault transfers slip from south-north to southeast-northwest trend as it passes this alluvial fan and produces some secondary fault
traces near the main fault scarp. The main fault strikes ~338° at the northwestern margin of this alluvial fan (Figures 3 and 7; Ganev et al., 2010).

Northwest of the TR alluvial fan there is an alluvial surface that exhibits no notable fault scarp, suggesting an age that is intermediate in age between the TR alluvial fan and the active channel, which has a well-defined easterly-northeasterly dip, as illustrated in the LiDAR DEM. We term this intermediate alluvial surface IAF (Figure 8c). The active channel has a narrow reach at the southwest end (black line in Figure 8e) and is characterized by a broad, anastomosing stream channel (ASC) reach within ~300 m of the fault on the southwest wall and continuing across the map area on the northeast wall. The ASC has split the active channel into a broad drainage system. Located between the TR alluvial fan and the active channel ASC on the northeast wall is a small (~100 × 300 m) area of this IAF surface that has escaped reworking by the active channel ASC (IAF*, outlined by the yellow dashed in Figure 8e). Paleostream channels across its surface have the same trend as the broader IAF surface. The TR-IAF* boundary likely formed at the same time as or slightly later than the formation of the broader IAF surface, i.e., some unknown amount of time after the formation of TR alluvial fan.

The multiple surfaces in close proximity to the TR alluvial fan suggest a complex erosion-deposition history, complicating the interpretation of fault displacement. Assuming the TR-IAF*/IAF* boundary had a straight shape at its formation we interpret a displacement between 90–200 m, depending on the time and degree of incision of the ASC (90 m if the ASC incised exclusively into the IAF, and the current TR-ASC boundary was a maximum extent of the IAF before incision of the ASC; 200 m if the ASC incised exclusively into the TR alluvial fan, and the current southwestern IAF-ASC boundary was the maximum extent of the TR alluvial fan before incision of the ASC. Figures 7i-7p, 8c-8g). Both of these interpretations assume that erosion has been equal along TR edge at both sides of the faults. Since erosion might not have been equal, due to differences in incision or changes in the anastomosing drainage patterns, alternative restorations are possible. Local geometric complexities due to fault steps (e.g., fault bending, vertical deformation, multiple fault strands), could also exert control on surface processes and mask reworking of features. For example, the edge of the TR alluvial fan may originally have had a more curved shape before the deposition of IAF or the incision of ASC. In that case, the displacement could be <90 m. Figures 7c-7h and 8b show a restoration of about 20 m displacement. We note that the TR alluvial fan edge immediately southwest of the fault is smoothed by erosion, and the TR alluvial fan edge immediately northeast of the fault may have been reworked by a creek that runs close to the fault, or be partially buried by colluvium. (Figures 8a and 8b).

Using the LaDiCaoz software, we derive a displacement estimate of 169 m by minimizing misfit between two elevation profiles along two sides of the fault trace (Figure 9). However, stream channels near the TR alluvial fan offset are not significantly deeper than surrounding surfaces, and elevations close to the fault may have been modified by desert flow of vertical motion. Thus, correlating elevation profiles with LaDiCaoz can result in an estimate with high uncertainty for this alluvial fan. The wide range of possible displacement estimates for the TR alluvial fan highlights the challenges in reconstructing the offset of a structurally complex and relatively young landform subjected to rapid reworking by desert flow.

4. New age estimates

Both the numerical dating and offset reconstruction can cause significant uncertainties in geologically-derived fault slip rate estimates. Ideally, the offset feature to be dated would have formed over a short interval of time, sometime after fault initiation. Alluvial fans in the southwestern US mainly represent Pleistocene and younger features, are thought to have formed in discrete intervals associated with climatic cycles/transition (e.g., McDonald et al., 2003; Dorn, 2009; Miller et al., 2010; Shepard et al., 2018), and hence can be useful for estimating geological fault slip rates.

We used several independent techniques and cross-correlated the results to estimate the ages of our two alluvial fans. The degree of rubification (Fe oxide coating) or desert varnish (Fe-Mn oxide coating) on surface clasts, the development of desert pavement, and the presence or absence of a well-developed caliche horizon, serve as qualitative age indicators. By these measures, the ALR alluvial fan is clearly older than the TR alluvial fan. Quantitative techniques such as terrestrial cosmogenic nuclides (TCN) and optically stimulated luminescence (OSL) dating were also used in this study. Both qualitative and quantitative techniques are described below.
4.1 Desert pavement development

Alluvial fan surfaces vary in the ratio of eolian fine sediments and stony pavement (desert pavement) with the percentage of stony pavement increasing with time. Wells et al. (1985) quantified this process for the northeast Mojave by estimating the percentage of stony pavement on a series of basalt flows of known age. Assuming similar processes apply to our area, the technique can be used to estimate a minimum age for a given alluvial fan surface (because of saturation effects the technique may not define an upper bound). Since the desert pavement is weak to moderately developed in the younger surfaces, we only analyzed the ALR alluvial fan surface. We estimated the density of stony pavement on the ALR alluvial fan surface based on surface color and correlation with exposed sand on 4 randomly selected photographs of the fan surface (Supplementary Figures S3-S6). Results suggest a minimum age of about 259 ka for the ALR alluvial fan (Figure 10a).

4.2 Carbonate rind thickness

Carbonate coatings on cobbles and boulders increase in thickness with increasing age. Amoroso (2006) developed a carbonate rind thickness model for the Mojave Desert, where rind thickness (in mm) is $0.0889 + 0.0079 \times$ (surface age in ka). In principle, this model could be used to estimate an age for the ALR alluvial fan, which has well developed coatings on cobbles and boulders (Supplementary Figure S2). However, a range of thicknesses is observed, leading to a rather wide range of age estimates, with a maximum of 482 ka for the ALR alluvial fan (Figure 10b). In addition, the model of Amoroso (2006) is only calibrated to ~130 ka, hence the accuracy for the ALR alluvial fan, which is likely much older, is not established.

4.3 Soil chronostratigraphy

Soil profile descriptions were performed in the field following standard techniques (Schoenenberger et al., 2012) (Supplementary Table S1). Profile development indices for observed profiles were calculated based on field observations following the method of Harden (1982). All of the profiles contained a ~10-cm-thick surface horizon with vesicular pores, secondary carbonates, and generally finer soil textures than underlying horizons indicating substantial contribution of fine grained eolian material to surface horizons (Wells et al., 1985).

The Calico-Pit2 profile in the TR alluvial fan exhibits minimal soil development in terms of pedogenic structure formation, reddening, and clay accumulation, with several lithologic discontinuities observed along with stratified sands and gravels in the subsurface. Subsurface horizons do contain secondary carbonates, mainly in the form of coatings on the bottom of gravels. The lack of pedogenic alteration resulted in a taxonomic classification of Typic Haplocambid (Soil Survey Staff, 2014), and a profile development index value <10 consistent with middle to early Holocene aged soils observed in other areas of the Mojave (Harden et al., 1991).

The CalicoA and Calico-Pit3 profiles in the ALR alluvial fan exhibit greater degrees of soil development, with substantial reddening and secondary carbonate accumulation, and a near completely interlocked surface pavement. The soils classify as Typic Haplocalcids (Soil Survey Staff, 2014), and profile development indices for these profiles ranged from ~18–25, with a greater degree of development in the CalicoA profile. These profile development indices are consistent with soils dated to ~200–250 ka in other Great Basin soil chronosequences (Harden et al., 1991). Field observation of the CalicoA profile suggested it may have been partially affected by erosion based on its soil morphologic features and location on the edge of the fan surface. Excavation of shallow pits in the center of the ALR alluvial fan bodies indicated a depth of 35–40 cm to reach the top of the Bk2 horizon versus a depth of 28 cm in the sampled and described CalicoA profile, suggesting a potential loss of 7–12 cm of depth due to erosion. Including this additional depth in the calculation of the profile development index pushes the age correlation estimate close to 300 ka.

4.4 Optically stimulated luminescence dating

OSL dating determines the time elapsed since a sediment sample was last exposed to daylight (Aitken,
1998). The method relies on the interaction of ionizing radiation with electrons in semi-conducting minerals within buried sediment, which results in metastable accumulation of charge. Illumination of the sediment releases the charge as a measurable emission of photons (luminescence). The methods assume that mineral grains during or immediately prior to the transport were exposed to daylight to set them to their geological zero residual level. Upon burial, day light exposure ceases and essentially the luminescence signal begins to accumulate due to the radiation arising from the decay of ambient radioisotopes that include U, Th, Rb and K, and from cosmic rays. Given that, as a first approximation, the radiation exposure (the dose rate $D_R$) is constant over the timescales of interest, luminescence builds up (equivalent dose $D_E$) in the minerals in proportion to the duration of burial and the concentration of the radioisotopes in the sample environment and the cosmic dose. The depositional age ($A$) of the sample is thus a ratio of luminescence acquired and the rate of luminescence acquisition, i.e., $A = D_E / D_R$ (Aitken, 1998; Murray and Olley, 2002; Singhvi and Porat, 2008).

Since the age of the ALR alluvial fan is likely to be beyond the range of applicability for the OSL technique, we only sampled the TR alluvial body, at the Calico-Pit2 (location shown in Figure 3b). Three OSL samples were collected at 75, 55 and 33 cm of depth (Calico F1, Calico F2 and Calico F3 respectively) by hammering 15-cm-long, 5-cm diameter plastic tubes into the sediment (Supplementary Figure S7). Detailed descriptions of the OSL sample processing steps are given in Supplementary Text S1.

Table 1 presents the radioisotope, water content, and cosmic dose, $D_R$, $D_E$ and OSL age for the samples. OSL characteristics and age determination are also discussed in more detail in Supplementary Text S1, Figures S8 and S9. The OSL ages range from $5.0\pm0.4$ ka (shallow sample) to $5.8\pm0.4$ ka (deep sample).

### 4.5 Terrestrial cosmogenic nuclide $^{10}$Be surface exposure dating

Cosmic rays generate TCNs in Earth’s atmosphere and surface that are produced at a known rate and can be used to date a variety of materials and processes. For alluvial fan surfaces, it is useful to focus on techniques where the radionuclides are produce in situ, a technique known as TCN surface exposure dating. Since quartz is a common rock-forming mineral rich in oxygen, a spallation reaction that transforms $^{16}$O to $^{10}$Be ($1.4\times10^5$ years) is especially useful (Gosse and Phillips, 2001). Comparison of surface samples to depth profiles can be helpful in establishing the degree of inheritance of $^{10}$Be, which otherwise can lead to anomalously old age estimates (alluvial fans often develop by re-mobilizing older alluvial material, and inheritance is a particular problem for younger fans, where the majority of the $^{10}$Be signal can be inherited). Overland sheet flow, erosion, and bioturbation can also disturb surface boulders, leading to anomalously young ages. Older alluvial fans are more likely to experience such disturbance.

We collected rock samples (>150 g each) from surface boulders and cobbles on the ALR and TR alluvial fan surfaces (red dots in Figure 3b). Sediment samples for depth profiles were collected at depth intervals of ~30 cm for three pits to a depth of 2 m: Calico A and Calico-Pit3 for the ALR alluvial fan, and Calico-Pit2 for the TR alluvial fan (light blue squares in Figure 3b). Supplementary Figure S10 shows pictures of collected samples. Both types of sampling were used for TCN dating using $^{10}$Be. Rock samples were chosen following criteria described in Gray et al. (2014), Frankel et al. (2015), and Hedrick et al. (2017), including: 1) large size, typically >50 cm in length; 2) stable boulders inset into the ground; 3) little sign of erosion; and, 4) quartz rich lithology. If boulders were absent whole cobbles were collected. The highest positions on alluvial fan surfaces were selected for depth profile trenches to minimize the possibility of surface erosion. All rock samples were prepared with standard procedures, including crushing, magnetic separation, heavy-liquid mineral separation (only for samples with significant heavy minerals or feldspars), etching, dissolution, purification, and target loading. Sediment samples went through the same processing procedures except for crushing: we sieved sand grain sizes between 0.25 and 0.5 mm before subjecting them to the procedures described above. Sand samples from Calico-Pit2 and Calico-Pit3 profiles did not yield a sufficient quantity of quartz, thus in order to increase the amount, small quartz-rich pebbles were added, with sizes between 0.5 and 12.5 mm; these pebbles were crushed, later sieved and added to the sand quartz fraction of 0.25 to 0.5 mm. Detailed descriptions of these samples and the TCN sample processing steps are given in Text S2 of the Supplement. Uncertainties for all ages are estimated at $2\sigma$ (95% confidence interval).

#### 4.5.1 Autumn Leaf Road alluvial fan

Ten rock samples on the ALR alluvial fan surface yield ages that range from ~75 to 346 ka (Figure 11 and Supplementary Table S2). Eight have ages of ~100 ka, while the other two (erosionally resistant quartz samples) have ages that are >300 ka.
To better constrain results and assess possible inheritance and erosion effects, depth profiles with samples collected at various depths from two trenches were used to help assess the validity of the surface exposure ages: one trench (CalicoA) at the southern corner of the main body of ALR alluvial fan, and a second one (Calico-Pit3) at the small fan surface to the northwest. Ideally, an exponential decrease of TCN concentration with depth is expected if the alluvial fan formed in a simple way, with no disturbance since formation and with all sediments that later became part of the alluvial fan containing the same concentration of inherited TCNs as that of the initial alluvial fan formation. The concentration of $^{10}Be$ versus depth profile thus contains information on exposure age, erosion history and inheritance. Figure 12(a) shows the $^{10}Be$ concentration versus depth profile for CalicoA, with the expected trend (decreasing $^{10}Be$ with depth).

A Bayesian-Monte Carlo simulation allows simultaneous estimation of age, erosion history, and inheritance (Hidy et al., 2010; Hidy, 2013). Loose constraints are applied, with age, erosion and inheritance allowed to vary between conservative high and low values, and Monte Carlo techniques are used to derive best estimates of the model parameters. The allowed ranges of the constraints used here are: 0–1200 ka for exposure age, 0–0.5 cm/ka for erosion rate, and 0–1.3×10$^9$ atoms/g for inheritance. No shielding was applied for the simulation (all samples were located far from high-relief features, and the environment is not currently favorable for snow cover). We assume an attenuation length of 164±5 cm and a stochastic uniform density of 1.9–2.5 g/cm$^3$, consistent with previous studies in similar environment (e.g., Hidy et al., 2010; Owen et al., 2011; and Gray et al., 2014). We then estimate the probability density functions for exposure age, erosion rate and inheritance (Figures 12a-12e). The most probable Bayesian age for CalicoA is 222 ka. However, without tight constraints on the other two parameters, the age estimate has a broad distribution, with a 2σ range between 196 and 332 ka (Fig. 12c, note tail on the upper bound is significantly longer). The 2σ upper limit of inheritance is 7.06×10$^4$ atoms/g, equivalent to an age of ~15 ka.

We applied the same method to the depth profile of Calico-Pit3, with constraint ranges as follows: 0–3000 ka for exposure age, 0–2.6 cm/ka for erosion rate, and 0–4.4×10$^9$ atoms/g for inheritance (a conservative high will produce unrealistically high inheritance, thus the maximum $^{10}Be$ concentration among the 6 samples was used as high-end constraint). The result gives a wide age distribution, with 2σ range between 99 ka and 2401 ka. Note that the most probable erosion rates from two depth profiles are smaller than the 3.05±0.62 cm/ka erosion rate estimated at the Calico Archaeological site, ~19 km away from our study area (Owen et al., 2011). Assuming the same erosion rate as Owen et al. (2011) yields a poor fit to the depth profiles and unrealistically old exposure ages. Due to the detailed soil profile analysis conducted, we recognize at CalicoA, that the upper section of the surface was eroded about 7–12 cm (section 4.3). Adjusting the depths for samples in the depth profiles by adding a 10 cm depth results in a slightly lower erosion rate that balances the depth adjustment, but the age and inheritance are essentially the same. While imperfect knowledge about erosion over the lifetime of the alluvial fan surface can limit the precision of the age estimates from depth profiles, these age limits allow us to exclude some rock samples from further study. For examples, rock samples with apparent ages of ~100 ka or less are clearly not representative, perhaps due to incomplete exposure (partial burial) or due to outer surfaces may have been eroded.

Two erosional resistant quartz samples yield apparent ages of >300 ka. We use the apparent age of the oldest sample (346±24 ka) as the most reliable estimate of the exposure age for the ALR alluvial fan surface. The depth profile results suggest that inheritance would not change this age significantly. This age estimate is consistent with results from the carbonate rind thickness, desert pavement density, and soil chronostatigraphy techniques (Figure 13a).

4.5.2 Troy Road alluvial fan

Among nine dated rock samples collected on the TR alluvial fan surface, one (Calico-6) has an age of 250 ka. This age is incompatible with the observed soil development here (no visually obvious rubification and carbonate coating). Inheritance from older units likely explains this anomalously old age. The remaining eight samples still exhibit a wide range of ages, from 10.9 to 70 ka (Figure 11). The depth profile for this fan, Calico-Pit2 does not show the expected exponential decrease in $^{10}Be$ concentrations with depth (Figure 12k), indicating inheritance saturation. The OSL samples provided ages in the range of 5.0±0.4 to 5.8±0.4 ka for the upper meter of the fan deposit, suggesting a Holocene age for this fan surface and indicating the fan sediments were deposited in a relatively short period of time. The young age and relatively fast sedimentation indicated by the OSL results, combined with inheritance saturation of $^{10}Be$ in depth profile, presumably explains the lack of exponential decrease with depth of the $^{10}Be$ concentrations.

For TCN dating techniques, understanding and quantifying inheritance is important. This is especially true for younger deposits and surfaces which can cannibalize older fan material, and have not
had sufficient time to generate a unique age signature. We estimated an inheritance of ~60 ka from our Calico-Pit2 depth profile. However, most of the rock samples from the TR alluvial fan have young apparent ages that are incompatible with inheritance saturation, indicating that inheritance in boulders and sand can vary significantly.

All of our TCN dated boulders are much older than the OSL results. There are two possibilities to explain the TCN dating of the TR alluvial fan: 1) The true exposure age of the TR alluvial fan is closest to the young cluster (10.9 ka to 17.1 ka) of apparent ages from rock samples with inheritance subtracted. The other rock samples have larger amounts of inheritance, and the depth profile represents an average inheritance for the fan material; 2) The true exposure age is significant older than the young cluster of apparent ages. In this case, a complex formation-exposure history has occurred, such that the depth profile does not show the expected exponential decrease in $^{10}$Be concentrations with depth. The first possibility agrees with the soil development and OSL dating results. The second possibility requires that the youngest four samples have experienced significant erosion or shielding after fan formation, which we think is unlikely given its relatively young age based on lack of rubification and desert pavement development.

We thus favor the first explanation. In this case, inheritance estimated from the depth profile (equal to ~60 ka) does not equal the inheritance of most of the rock samples (~45 ka). Since the scatter of apparent ages is largely determined by inheritance, we therefore discard the older apparent ages. Based on the four boulders with youngest apparent ages we estimate the weighted mean and 2σ, and obtain 14.0±5.8 ka. Note this is a maximum age estimate for the TR alluvial fan as it contains inheritance that is unknown. If the OSL dated age of ~5 ka represents the true age, then inheritance accounts for ~9 ka among the youngest cluster of ages. Dating sediments in the Mojave through OSL can be also challenging. OSL measurements may give ages that are too young, for example due to low OSL sensitivity, poor quartz characteristics, high dose rates, and low water estimates (e.g. Owen et al., 2007; Lawson et al., 2012). We thus assume 5 ka as a minimum age for the TR alluvial fan.

These numerical dating results place the age of the TR alluvial fan in the range 5±0.4 to 14.0±5.8 ka, consistent with proposed ages for alluvial fan generation by Miller et al. (2010), and consistent with soil chronology at this site.

5. Slip rate estimates

For the ALR alluvial fan, the displacement of 1110±110 m and the TCN exposure age from the sample with the largest apparent age (346±24 ka) yields a slip rate of 3.2±0.4 mm/yr. For the TR alluvial fan, the OSL dating yields a minimum age of 5.0±0.4 ka, and the TCN surface exposure dating gives a maximum estimate of 14.0±5.8 ka. Displacement estimates of the TR alluvial fan are highly uncertain, hence the slip rate is not well constrained by our data (Figure 14). While a displacement estimate of 20 m, and the OSL age of 5 ka define a slip rate similar to the ALR alluvial fan data, the complex offset geometry and spread of geochronology data for this alluvial fan allow alternate interpretations. For example, the displacement estimate of 90 m and TCN exposure age of 14.0±5.8 ka define a faster slip rate, 6.4±2.7 mm/yr; the same displacement but using the OSL dated age results in an even higher slip rate. In the discussion below, we use the slip rate estimate based on data from the ALR alluvial fan as the best estimate for the long-term average (several hundred thousand year) slip rate of the Calico Fault.

6. Discussion

Our new slip rate estimate of 3.2±0.4 mm/yr based on the ALR alluvial fan data is considerably faster than previously published values for the Calico Fault (Table 2). This new slip rate is faster than the estimate of 1.8±0.3 mm/yr from the 56±7.7 ka old ‘K’ alluvial fan of Oskin et al. (2007a), more than double the slip rate estimate of 1.4±0.4 mm/yr from the 650±100 ka old ‘B’ alluvial fan by Oskin et al. (2007a), and more than double the 1.4 ±/−0.4 mm/yr estimate from a 17.1 ±/−1.6 ka ‘QB’ unit southwest of the Rodman Mountains reported by Selander (2015). Although our estimate may represent a maximum slip rate due to limited knowledge of the erosion history and corresponding uncertainty in the alluvial fan age (i.e., the alluvial fan could be older), the lower slip rate limit should not be much lower, for three reasons: 1) the probability density function (pdf) of the age–depth profile CalicoA (Figure 12c) skews strongly to ~300 ka, with a correspondingly small probability for ages >600 ka; 2) a slip rate of 1.8 mm/yr for the ALR alluvial fan requires that its exposure age would be >600 ka given the 1110 m displacement, incompatible with the carbonate rind thickness and soil development data, which require an age much younger than this; and 3) our slip rate estimate for the TR alluvial fan, while it has a larger uncertainty, is also consistent with a faster rate (though it does not require it; Figure 14). For example,
assuming the smallest possible displacement (20 m) and the OSL results (5.0±0.4 ka) gives a rate of ~4 mm/yr. Any of the larger displacement estimates requires a correspondingly faster rate if the OSL age estimate is correct.

Since the various published rate estimates and our own estimates are based on offsets from different locations along the Calico Fault and features with a range of ages, there are several possible explanations for the differences, including: 1) the slip rate on the Calico Fault changes along strike; 2) the slip rate on the Calico Fault changes with time; 3) one or more of the assumptions used to estimate the slip rates are in error; and/or 4) all of the data are correct in a strict sense but the estimates represent on-fault slip rates at the corresponding sites. In this case the fastest one may be closest to the true slip rate due to different amounts of slip localization and off-fault deformation.

Regarding a change in slip rate along strike (our first explanation), Oskin et al. (2007a) and Selander (2015) suggest that slip in the Calico-Blackwater Fault system varies spatially, with slip transferring from the Calico Fault to the Harper Lake and Blackwater faults via a set of thrust ramps or absorbed by folding adjacent to the Calico Fault. The two studies referenced above are separated by only ~8 km, and have rates that differ by ~30%. Our study area is ~7 km from the study area of Oskin et al. (2007a), and ~15 km from the study area of Selander (2015) (Figure 15). To support an increase from 1.4 or 1.8 to 3.2 mm/yr would require that nearby faults (the fastest one is Camp Rock Fault, with slip rate ≤1.4 mm/yr (Oskin et al. 2008)) transfer almost all their slip to the Calico Fault within 15 km. While it seems overly complicated, we cannot preclude this possibility considering the highly discontinuous and complex nature of the Mojave section of the ECZS. More studies are needed to explore the possibility of slip rate transfer along strike, or communication among nearby faults.

As for a change in slip rate over time (our second explanation), the displacement of the ALR alluvial fan (110±110 m) is slightly larger than the ‘B’ alluvial fan (900±200 m, with a surface exposure age estimate of 650±100 ka) of Oskin et al. (2007). To reconcile the displacement and age data, the Calico Fault would need to be inactive between the formation time of the ‘B’ alluvial fan and the formation time of the ALR alluvial fan. Unless the ALR alluvial fan has an exposure age similar as the ‘B’ fan, this seems unlikely, based on the age constraints described above.

We now consider the possibility that different rate estimates may be caused by incorrect assumptions used in the slip rate calculations (our third explanation). For example, offset reconstructions use landforms such as alluvial fans, but erosion may blur key features. In the case of the TR alluvial fan, the degree of incision by the anastomosing stream channels into the TR alluvial fan can change the displacement and slip rate estimate by more than a factor of two. Since displacement is in the numerator for slip rate estimates, uncertainty in the displacement estimate has more of an effect on the rate estimate for younger alluvial fans. However, older alluvial fans may suffer more erosion, making it difficult to estimate accurate displacements for these features. Given the age ranges for the TR alluvial fan offset from Oskin et al. and TCN dating (5.0–14.0 ka), and a plausible maximum range of slip rates for the Calico Fault (1.4–12.0 mm/yr; Sauber et al., 1994; Oskin et al., 2007a; Selander, 2015; McGill et al., 2015; this study), plausible displacements span a large range, from 7–168 m (Figure 14). If correct, this suggests that the ASC surface may not be incised exclusively into the TR alluvial fan.

Age determinations can also cause large uncertainties in slip rate estimates. TCN exposure dating techniques often result in large uncertainties due to a variety of geologic factors (Owen et al., 2011), requiring data editing that could introduce systematic biases. While such editing is usually based on sound geological criteria, e.g., relative ages derived from field observations, these assignments become more difficult for older fans. Wells et al. (1985) and Oskin et al. (2007a) note that criteria such as surface morphology and clast weathering tend to approach steady state with increasing age. Also, as alluvial fans get older, the possibility of surface disturbance increases (Owen et al., 2014). Northern hemisphere alluvial fans older than 300 ka have experienced three complete glacier-interglacial cycles, increasing the likelihood of surface disturbance by erosion and occasional minor deposition during wetter and cooler climate periods. This could result in ages that are too young.

While the 10Be TCN technique used here has been widely employed for surface exposure dating, it does rely on several assumptions for both pre- and post-formation history, including inheritance, erosion, and shielding. For the ALR alluvial fan, although most rock samples have age estimates of ~100 ka, we interpret these as underestimates of the true age of the landform because these ages are incompatible with soil, pavement, and carbonate rind development, and are well beyond the age estimate from the depth profile CalicoA. There are several explanations for ages that are too young, including shielding from cosmic rays due to sediment cover and later exhumation, toppling or rotation of the sampled clasts during erosional events, weathering (spallation), and bioturbation. The large range of age estimates (both rock samples and depth profiles) in both our study and other published studies reflects the multiple different surface processes operating on alluvial fan surfaces and clasts that can make 10Be TCN data and other exposure age data challenging to interpret. We note, however, that our data exhibit no more scatter than other comparable studies (Figure 16).

Limited sampling can also lead to uncertainties and biases in geological slip rate estimates. Ideally, hundreds of samples would allow more rigorous assessment and statistical characterization of the
various processes listed above that affect results. Unfortunately, this is well beyond current capabilities. For example, our preferred age estimate for the ALR alluvial fan is based on only the oldest rock sample; we interpret our other samples to be biased to younger ages. Similarly, in the study of Oskin et al. (2007a), the ‘B’ fan was assigned an age of 650±100 ka based on a single sample, the one with the oldest cosmogenic 3He exposure age (653.4±19.8 ka; the second oldest age is 418.9±12.6 ka), plus 40Ar/39Ar dating of two flows with ages of 770±40 ka and 735±9 ka, and all other 3He dated samples yield much younger ages. The samples used for 40Ar/39Ar dating were also located several kilometers away from the source (Pipkin Basalt Flows, see Figure 2A in Oskin et al. (2007a)). The ages of these samples may therefore represent upper bound for the formation age of the ‘B’ fan. In the same study, the ‘K’ fan was dated using both TCN 3He and 10Be. While 10Be dating gave consistent ages, 3He dating results spanned a large range (more than a factor of 2) and hence were not used to define the exposure age of the ‘K’ fan. In both our study and the study of Oskin et al. (2007a) (and indeed most similar studies) the limited number of dated samples, the wide range of apparent ages, and the necessity of using selection criteria, increase the chances of biased results (Figure 16).

Young alluvial fans likely experienced less erosion but inheritance in TCN dating can be more significant due to their short exposure periods, as illustrated by the OSL results for the TR alluvial fan. Old alluvial fans likely experienced more erosion, but inheritance is dwarfed by the long post-formation exposure ages. Based on our data and several other studies (e.g., Oskin et al. 2017a; Frankel et al., 2011), the apparent ages of rock samples from an alluvial fan surface dated by TCN exposure age techniques tend to be distributed similarly to a chi-square distribution (Figure 17). For a young alluvial fan (a few tens of ka or younger; e.g., ‘K’ fan and the TR alluvial fan in Figure 16), apparent ages tend to skew towards older values because of inheritance (e.g., this study; Oskin et al. 2007a; Frankel et al., 2011). Hence the OSL results may be more reliable. In contrast, for older alluvial fans (hundreds ka or older; e.g., ‘B’ fan and the ALR alluvial fan in Figure 16), apparent ages tend to skew towards younger values because of disturbance and erosion (e.g., this study; Oskin et al., 2007a). While these considerations help in selecting reliable ages from an ensemble of apparent ages, it is clear that biases can occur.

These considerations suggest to us that the uncertainties of geologic slip rate estimate are often under-estimated (see also Bird, 2007, and Zecchin and Frankel, 2009). Even judicious selection criteria can lead to significant scatter in results (Figure 18). If Bird’s (2007) criterion for the minimum number of independent estimates required for reliable rate determinations is applied, even with our current study, we are perhaps having a robust picture for the slip rate history of individual faults in the Mojave section of the ECSZ, or the summed geologic rate across the shear zone.

Regarding on-fault slip rates (our fourth explanation), the slip rate of 3.2±0.4 mm/yr from the ALR alluvial fan data is significantly faster than the estimate of 1.8±0.3 by Oskin et al. (2007a) at a site that is only ~7 km away. However, these data could be reconciled if: 1) fault slip is almost completely localized onto the surface offset at ALR alluvial fan during its slip period; and 2) the offset studied by Oskin et al. (2007a) missed some off-fault deformation.

Dolan and Haravitch (2014) considered all faults in the Mojave section of the ECSZ as structurally immature, with strain not yet completely localized onto a narrow high strain fault core. Using their criteria, most ECSZ fault studies have underestimated slip rates. Dolan and Haravitch (2014) suggested that the slip rates of immature faults in the Mojave ECSZ have been underestimated by ~40%, even along straight, continuous, structurally simple sections of surface rupture. Using modified fault configurations, Herbert et al. (2014a) found that off-fault deformation accounts for 40±23% of the total strain across the ECSZ, with higher percentages near places where faults terminate or bend. If the value of 40±23% applies to most major faults in the Mojave region including the Calico Fault, then scaling Oskin et al.’s (2007a) result by this amount gives 3.0±1.3 mm/yr, equivalent to our result within uncertainties. If this explanation is correct, it implies that surface geomorphic markers in this region record a variable portion of total displacement. In a similar environment, Fletcher et al. (2014) mapped surface ruptures caused by the Mw 7.2 El Mayor-Cucapah earthquake (2010), finding that surface displacements varied along strike by orders of magnitude within a few kilometers along the trace of the fault.

It should be noted that Oskin et al. (2007a) considered the possibility of off-fault deformation within a few hundred meters of the Calico fault. However, a related factor not considered by Oskin et al (2007a) is the possibility of an unexposed and unmapped fault strand farther away that locally carries some of the slip at the more northern Calico site, but is not present at the location of the ALR and TR alluvial fans. Such unmapped strands may be especially problematic in the Mojave Desert, where widespread young alluvial deposits obscure the geomorphic effects of slow-moving strike slip faults and limit the number of clear offset markers. For example, Rockwell et al. (2000) noted that the southern part of the Landers earthquake rupture occurred on a previously unmapped fault.

The total displacement on a fault can help to constrain estimates of present day fault slip rate if the initiation age of the fault is known and if the slip rate history follows a simple evolutionary path, e.g., a monotonic increase in rate through time until fault maturity is reached (Gourmelen et al., 2011). Total displacement may also help to address the issue of unaccounted off-fault deformation. Assuming it is
estimated using an older, well-defined offset marker, total displacement should account for both on- and off-fault deformation. More generally, total displacement and slip rate history together help to define the evolutionary process of faulting and fault maturity.

Total displacement across the Calico Fault is estimated to be 9.8 km (Glazner et al., 2000; Oskin et al., 2007a). While we lack hard data on the age of slip initiation, some constraints are available. The most likely time for initiation of ECSZ faulting as a whole is sometime after the inland jump of the Pacific-North America plate boundary. Marine incursion into the northern Gulf of California is dated at 6.2±0.2 Ma (Bennett et al., 2015). The ECSZ likely formed soon after this time. Lee et al. (2009) suggested a 2.8 Ma initiation age for the Saline Valley–Hunter Mountain–Panamint Valley fault system north of the Garlock Fault, which may be kinematically related to the central faults of the Mojave ECSZ, including the Calico Fault. Dixon and Xie (2018) proposed a 4.1 Ma initiation age for the Calico Fault. Andrew and Walker (2017) proposed an initiation age for the Blackwater Fault, immediately northwest of the Calico Fault, at or after 3.8 Ma. In the analysis that follows, we investigate model with initiation ages between 2.8 and 6.2 Ma.

While firm conclusions cannot be drawn regarding slip rate and its relation to initiation age and total displacement, it is possible to rule out certain combinations of parameters. For example, both constant rate and constant acceleration models are inconsistent with initiation ages of 4.1 Ma or younger and present-day slip rate of 1.8 mm/yr (the total displacement would be smaller than observed; Figure 19a).

Gourmelen et al. (2011) proposed a damage growth model for young faults that results in an intermediate style of fault evolution, more rapidly than a constant acceleration model, but more slowly than a constant rate model (the latter implies essentially instantaneous acceleration at fault initiation). In the damage growth model, slip rate ramps up on time scales of several hundred thousand to several million years from zero at fault initiation to some steady state rate. The time from fault initiation to maximum acceleration is given by the Rayleigh scale parameter $S$, which occurs at roughly one half the final rate (Gourmelen et al., 2011). Figure 19b shows slip rate as a function of time for this model for a fault with 9.8 km displacement and an initiation age of 4.1 Ma, i.e., slightly older than the age proposed by Andrew and Walker (2017). Possible present-day rates range from 2.5 to 5.6 mm/yr. Figure 19c shows the range of permissible models for all initiation ages ($t$) between 2.8 and 6.2 Ma, using a range of values for $S$ between 0.1 to 2.5 Ma. For a 2.8 Ma initiation age, possible values of present-day slip range from 3.5–9.3 mm/yr; For a 6.2 Ma initiation age, possible values of present-day slip range from 1.6–3.0 mm/yr.

If our new estimate of 3.2±0.4 mm/yr is representative of the true slip rate of the Calico Fault (explanation number 4), it increases the cumulative geologic slip rate for the Mojave section of the ECSZ. If the ‘slip deficit’ noted in previous studies is largely due to fault immaturity, off-fault deformation, and underestimated slip, we can better approximate the overall geologic slip across the ECSZ by using our new result for the Calico Fault, and scaling on-fault geologic slip rates for remaining five major faults by some amount of off-fault deformation. There are various ways to do this. Using Herbert et al.’s (2014a) estimate of off-fault deformation (40±23%) for the remaining five ECSZ faults, then the total geologic slip rate across the ECSZ becomes 10.9±3.1 mm/yr (using the rates in Oskin et al. (2008) for the other five ECSZ faults; uncertainty in off-fault deformation of Herbert et al. (2014a) is considered), equivalent within uncertainties to the geodetically-derived rate (Figure 20).

Given the importance of fault slip rate in seismic hazard estimates and fault evolution studies, the above discussion highlights the importance of continued research, and development of new approaches to fault slip rate determination over different time spans. For young alluvial fans such as the TR fan, perhaps the use of landscape evolution models convolved with active faulting could better account for the effects of erosion and refine the displacement estimates, and hence the slip rate estimates (Fig. 14).

We suggest that available data do not support a discrepancy between the summed geodetic rate across this section of the ECSZ and corresponding geologic rates averaged over the last few hundred thousand years, once the true uncertainty of these approaches has been considered. Since seismic hazard is closely related to fault slip rate, if discrepancies are observed, the faster rates (in this case based on the geodetic data) should be considered for seismic hazard estimates, until proven otherwise. This finding applies not only to the Mojave region, but also to other new or rapidly evolving plate boundary zones, where surface faults are immature, may be poorly exposed at the surface, and may have very low (or even zero) geologic slip rate estimates. The 2003 Bam earthquake in Iran, e.g., was responsible for ~30,000 fatalities, and occurred on a fault with limited surface exposure (Talebian et al., 2004; Fialko et al., 2005). Geologic studies on this fault, had they been done, would have indicated a slip rate of essentially zero. The tectonics of that region have clear similarities with the region studied here (Dixon and Xie, 2018).
7. Conclusions

Our new slip rate estimate for the Calico Fault of 3.2±0.4 mm/yr highlights the possibility that some fault slip rates in the Mojave Desert may have been under-estimated. This may be related to the limited number of studies in the region and the highly discontinuous and complex nature of the fault system. Among these immature young faults, where distributed deformation is common, total offset is not necessarily manifested as simple surface offset across a discrete fault plane; unmapped fault strands may be common. We also suggest that geologically determined slip rates have uncertainties that may be under-estimated. The overall ~10–12 mm/yr geodetic slip rate across the ECSZ is likely equivalent to the summed geologic rate across the region if our new estimate better represents the true slip rate along the Calico Fault system, if reasonable amounts of distributed deformation and unmapped faulting away from the major faults are taken into account (Figure 20), and if realistic uncertainties are considered. Until the origins of the apparent rate discrepancy for the ECSZ are fully understood, the faster geodetic rate should be considered for seismic hazard estimates in the region.

Acknowledgments: This research was supported by USGS-NEHRP grant G16AP00102 to PHW and THD, and G16AP00103 to LAO. We acknowledge Amelia Nachbar of the University of South Florida for help with the field work, and Antonio Luna for help with some sample preparation. Sarah Hammer and Kat Rivers at the University of Cincinnati are also thanked for help in sample preparation. We thank OpenTopography for providing LiDAR raster data, and USGS for providing high resolution aerial ortho-imagery. Mike Oskin, David Katopody and an anonymous reviewer are thanked for their thoughtful comments.

Disclosure statement: No potential conflict of interest was reported by the authors

References


Herbert, J.W., Cooke, M.L., Oskin, M., and Difo, O., 2014a, How much can off-fault deformation contribute to the slip rate discrepancy within the eastern California shear zone?. Geology, v. 42(1), p. 71-75. doi:10.1130/G34738.1


Hidy, A.J., Gosse, J.C., Pederson, J.L., Mattern, J.P. and Finkel, R.C., 2010, A geologically constrained Monte Carlo approach to modeling exposure ages from profiles of cosmogenic nuclides: An
example from Lees Ferry, Arizona. Geochemistry, Geophysics, Geosystems, v. 11(9). doi: 10.1029/2010GC003084


Rockwell, T.K., Lindvall, S., Herzberg, M., Murbach, D., Dawson, T., and Berger, G., 2000, Paleoseismology of the Johnson Valley, Kickapoo, and Homestead Valley faults: Clustering of


### Table 1 Summary of OSL dating results from extracted from sediment, sample locations, radioisotopes concentrations, moisture contents, total dose-rates, $D_E$ estimates and optical ages.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Location ($\pm$N/W)</th>
<th>Altitude (m)</th>
<th>Depth (cm)</th>
<th>U ($ppm$)</th>
<th>Th ($ppm$)</th>
<th>K (%)</th>
<th>Rb ($ppm$)</th>
<th>Cosmic $^b$, $^c$ (Gy/ka)</th>
<th>Dose-rate $^b$, $^d$ (Gy/ka)</th>
<th>Dispersity</th>
<th>Average equivalent dose $^g$, $^h$ (Gy)</th>
<th>Average equivalent dose $^g$, $^i$ (Gy)</th>
<th>OSL Age $^g$, $^h$ (ka)</th>
<th>OSL Age $^i$ (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calico</td>
<td>34.7925/116.63</td>
<td>596</td>
<td>75</td>
<td>2.5</td>
<td>14.5</td>
<td>100</td>
<td>0.21</td>
<td>4.395 ± 0.2</td>
<td>24 ± 15</td>
<td>19</td>
<td>23.4 ± 0.1</td>
<td>28 ± 0.1</td>
<td>25.42 ± 1.07</td>
<td>25.46 ± 1.07</td>
</tr>
<tr>
<td>F1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calico</td>
<td>34.7925/116.63</td>
<td>596</td>
<td>55</td>
<td>2.1</td>
<td>14.1</td>
<td>118</td>
<td>0.22</td>
<td>4.447 ± 0.2</td>
<td>24 ± 13</td>
<td>21</td>
<td>27.66 ± 0.1</td>
<td>32.08 ± 0.1</td>
<td>23.27 ± 2.1</td>
<td>5.2 ± 0.2</td>
</tr>
<tr>
<td>F2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calico</td>
<td>34.7925/116.63</td>
<td>596</td>
<td>33</td>
<td>2.3</td>
<td>14.9</td>
<td>3.1</td>
<td>0.22</td>
<td>4.731 ± 0.2</td>
<td>32 ± 15</td>
<td>40</td>
<td>29.91 ± 0.1</td>
<td>33 ± 0.1</td>
<td>23.84 ± 1.03</td>
<td>5.0 ± 0.1</td>
</tr>
<tr>
<td>F1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*a* Elemental concentrations from NAA of whole sediment measured at Activation Laboratories Limited Ancaster, Ontario Canada.

*b* Estimated fractional day water content for whole sediment is taken as 10% and with an uncertainty of ±5%.

*c* Estimated contribution to dose-rate from cosmic rays calculated according to Prescott and Hutton (1994). Uncertainty taken as ±10%.

*d* Total dose-rate from beta, gamma and cosmic components. Beta attenuation factors for U, Th and K compositions incorporating grain size factors from Mejdahl (1979). Beta attenuation factor for Rb is taken as 0.75 (cf. Adamiec and Aitken, 1998). Factors utilized to convert elemental concentrations to beta and gamma dose-rates from Adamiec and Aitken (1998) and beta and gamma components attenuated for moisture content. Dose-rates calculated through Aberystwyth University DRAC calculator (Durcan et al., 2015).

*e* Total number of single-aliquot measured.

$f$ Number of replicated equivalent dose ($D_E$) successfully measured determined from replicated single-aliquot regenerative-dose method (SAR; Murray and Wintle, 2000). These are based on recuperation error of < 10%.

$g$ Weighted mean and standard error equivalent dose ($D_E$) for all aliquots. The uncertainty includes an uncertainty from beta source estimated of ±5%.

$h$ Uncertainty incorporate all random and systematic errors, including dose rates errors and uncertainty for the $D_E$.

$i$ Mean and standard error equivalent dose ($D_E$) for minimum peak. If $D_E$ dispersion was >20%, then a 2 mixing model was considered. The uncertainty includes an uncertainty from beta source estimated of ±5%.
Table 2 Displacement, age and slip estimates from Oskin et al. (2007a, b), Selander (2015) and this study.

<table>
<thead>
<tr>
<th>Surface</th>
<th>Sample</th>
<th>DATING technique</th>
<th>Age (ka)</th>
<th>Age uncertainty (ka, 2σ)</th>
<th>Preferred rate * (mm/yr, 2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ALR (this study)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Displacement (2σ)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1110±110 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calico-9</td>
<td>Be</td>
<td></td>
<td>345.7</td>
<td>24.3</td>
<td></td>
</tr>
<tr>
<td>Calico-11</td>
<td>Be</td>
<td></td>
<td>331.7</td>
<td>20.4</td>
<td></td>
</tr>
<tr>
<td>Calico-12</td>
<td>Be</td>
<td></td>
<td>75.3</td>
<td>8.3</td>
<td></td>
</tr>
<tr>
<td>CA-14</td>
<td>Be</td>
<td></td>
<td>113.6</td>
<td>7.5</td>
<td></td>
</tr>
<tr>
<td>Calico-20</td>
<td>Be</td>
<td></td>
<td>112.1</td>
<td>25.4</td>
<td></td>
</tr>
<tr>
<td>Calico-23</td>
<td>Be</td>
<td></td>
<td>81.5</td>
<td>23.2</td>
<td></td>
</tr>
<tr>
<td>Calico-25</td>
<td>Be</td>
<td></td>
<td>107</td>
<td>19.5</td>
<td></td>
</tr>
<tr>
<td>CA-104</td>
<td>Be</td>
<td></td>
<td>70.4</td>
<td>9.0</td>
<td></td>
</tr>
<tr>
<td>CA-106</td>
<td>Be</td>
<td></td>
<td>106.2</td>
<td>7.2</td>
<td></td>
</tr>
<tr>
<td>CA-107</td>
<td>Be</td>
<td></td>
<td>82.8</td>
<td>13.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>96.5</td>
<td>7.5</td>
<td></td>
</tr>
<tr>
<td>TR (this study)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Displacement 90 m (note there are two possibilities, see Figures 7-9, 14)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calico-1</td>
<td>Be</td>
<td></td>
<td>41.4</td>
<td>3.3</td>
<td></td>
</tr>
<tr>
<td>Calico-2</td>
<td>Be</td>
<td></td>
<td>42.5</td>
<td>7.8</td>
<td></td>
</tr>
<tr>
<td>CA-102</td>
<td>Be</td>
<td></td>
<td>59.6</td>
<td>5.1</td>
<td></td>
</tr>
<tr>
<td>Calico -5</td>
<td>Be</td>
<td></td>
<td>70.4</td>
<td>9.9</td>
<td></td>
</tr>
<tr>
<td>Calico-6</td>
<td>Be</td>
<td></td>
<td>17.1</td>
<td>2.3</td>
<td></td>
</tr>
<tr>
<td>Calico-7</td>
<td>Be</td>
<td></td>
<td>12.1</td>
<td>2.2</td>
<td></td>
</tr>
<tr>
<td>Calico-8</td>
<td>Be</td>
<td></td>
<td>15.9</td>
<td>3.9</td>
<td></td>
</tr>
<tr>
<td>Calico-9</td>
<td>Be</td>
<td></td>
<td>249.7</td>
<td>24.9</td>
<td></td>
</tr>
<tr>
<td>Calico-10</td>
<td>Be</td>
<td></td>
<td>10.9</td>
<td>2.8</td>
<td></td>
</tr>
<tr>
<td>Calico-F</td>
<td>OSL</td>
<td></td>
<td>5.8</td>
<td>0.4</td>
<td></td>
</tr>
<tr>
<td>Calico-F</td>
<td>OSL</td>
<td></td>
<td>5.2</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td>Calico-F</td>
<td>OSL</td>
<td></td>
<td>5.0</td>
<td>0.4</td>
<td></td>
</tr>
<tr>
<td>SWRM (Selander, 2015)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Displacement (2σ)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>24±2 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Q2c: depth profile</td>
<td>Be</td>
<td></td>
<td>17.1</td>
<td>1.8±0.3</td>
<td>1.4±0.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B (Oskin et al., 2007)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Displacement (2σ)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>900±200 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C04-002-A</td>
<td>Ar/39</td>
<td></td>
<td>18966</td>
<td>4246</td>
<td></td>
</tr>
<tr>
<td>C04-002-B</td>
<td>Ar/39</td>
<td></td>
<td>906</td>
<td>46</td>
<td></td>
</tr>
<tr>
<td>C04-002-C</td>
<td>Ar/39</td>
<td></td>
<td>778</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>C04-002-D</td>
<td>Ar/39</td>
<td></td>
<td>759</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>C04-002-E</td>
<td>Ar/39</td>
<td></td>
<td>722</td>
<td>28</td>
<td></td>
</tr>
<tr>
<td>C04-002-F</td>
<td>Ar/39</td>
<td></td>
<td>2234</td>
<td>676</td>
<td></td>
</tr>
<tr>
<td>C04-002-G</td>
<td>Ar/39</td>
<td></td>
<td>907</td>
<td>116</td>
<td></td>
</tr>
<tr>
<td>C04-002-H</td>
<td>He</td>
<td></td>
<td>776</td>
<td>46</td>
<td></td>
</tr>
<tr>
<td>C04-002-J</td>
<td>He</td>
<td></td>
<td>756</td>
<td>18</td>
<td></td>
</tr>
<tr>
<td>C04-002-K</td>
<td>He</td>
<td></td>
<td>763</td>
<td>16</td>
<td></td>
</tr>
<tr>
<td>C04-002-M</td>
<td>He</td>
<td></td>
<td>777</td>
<td>52</td>
<td></td>
</tr>
<tr>
<td>C04-002-N</td>
<td>He</td>
<td></td>
<td>4044</td>
<td>598</td>
<td></td>
</tr>
<tr>
<td>C04-002-O</td>
<td>He</td>
<td></td>
<td>653.4</td>
<td>19.8</td>
<td></td>
</tr>
<tr>
<td>C04-002-P</td>
<td>He</td>
<td></td>
<td>218.7</td>
<td>6.7</td>
<td></td>
</tr>
<tr>
<td>C04-002-Q</td>
<td>He</td>
<td></td>
<td>418.9</td>
<td>12.6</td>
<td></td>
</tr>
<tr>
<td>C04-002-R</td>
<td>He</td>
<td></td>
<td>144</td>
<td>4.4</td>
<td></td>
</tr>
<tr>
<td>C04-002-S</td>
<td>He</td>
<td></td>
<td>73.5</td>
<td>1.9</td>
<td></td>
</tr>
<tr>
<td>C04-002-T</td>
<td>He</td>
<td></td>
<td>59.4</td>
<td>1.6</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C04-005</td>
<td>He</td>
<td></td>
<td>130.2</td>
<td>4.1</td>
<td></td>
</tr>
<tr>
<td>C04-006</td>
<td>He</td>
<td></td>
<td>106.8</td>
<td>3.2</td>
<td></td>
</tr>
<tr>
<td>C04-007</td>
<td>He</td>
<td></td>
<td>106.8</td>
<td>3.2</td>
<td></td>
</tr>
<tr>
<td>C04-009</td>
<td>He</td>
<td></td>
<td>155.5</td>
<td>4.7</td>
<td></td>
</tr>
<tr>
<td>C04-010</td>
<td>He</td>
<td></td>
<td>268.5</td>
<td>8.1</td>
<td></td>
</tr>
<tr>
<td>C04-015</td>
<td>He</td>
<td></td>
<td>95.0</td>
<td>3.2</td>
<td></td>
</tr>
<tr>
<td>C04-018</td>
<td>He</td>
<td></td>
<td>137.9</td>
<td>4.6</td>
<td></td>
</tr>
<tr>
<td>C04-011</td>
<td>He</td>
<td></td>
<td>57.1</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td>C04-012</td>
<td>He</td>
<td></td>
<td>63.3</td>
<td>1.7</td>
<td></td>
</tr>
<tr>
<td>C04-013</td>
<td>He</td>
<td></td>
<td>57.6</td>
<td>2.0</td>
<td></td>
</tr>
<tr>
<td>C04-014</td>
<td>He</td>
<td></td>
<td>55.9</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td>C04-016</td>
<td>He</td>
<td></td>
<td>53.4</td>
<td>1.4</td>
<td></td>
</tr>
<tr>
<td>C04-017</td>
<td>He</td>
<td></td>
<td>56.3</td>
<td>2.0</td>
<td></td>
</tr>
<tr>
<td>C04-018</td>
<td>He</td>
<td></td>
<td>48.3</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td>C04-020</td>
<td>He</td>
<td></td>
<td>63.3</td>
<td>1.9</td>
<td></td>
</tr>
<tr>
<td>K (Oskin et al., 2007)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Displacement (2σ)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>100±10 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C04-005</td>
<td>He</td>
<td></td>
<td>130.2</td>
<td>4.1</td>
<td></td>
</tr>
<tr>
<td>C04-006</td>
<td>He</td>
<td></td>
<td>106.8</td>
<td>3.2</td>
<td></td>
</tr>
<tr>
<td>C04-007</td>
<td>He</td>
<td></td>
<td>106.8</td>
<td>3.2</td>
<td></td>
</tr>
<tr>
<td>C04-009</td>
<td>He</td>
<td></td>
<td>155.5</td>
<td>4.7</td>
<td></td>
</tr>
<tr>
<td>C04-010</td>
<td>He</td>
<td></td>
<td>268.5</td>
<td>8.1</td>
<td></td>
</tr>
<tr>
<td>C04-015</td>
<td>He</td>
<td></td>
<td>95.0</td>
<td>3.2</td>
<td></td>
</tr>
<tr>
<td>C04-018</td>
<td>He</td>
<td></td>
<td>137.9</td>
<td>4.6</td>
<td></td>
</tr>
<tr>
<td>C04-011</td>
<td>He</td>
<td></td>
<td>57.1</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td>C04-012</td>
<td>He</td>
<td></td>
<td>63.3</td>
<td>1.7</td>
<td></td>
</tr>
<tr>
<td>C04-013</td>
<td>He</td>
<td></td>
<td>57.6</td>
<td>2.0</td>
<td></td>
</tr>
<tr>
<td>C04-014</td>
<td>He</td>
<td></td>
<td>55.9</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td>C04-016</td>
<td>He</td>
<td></td>
<td>53.4</td>
<td>1.4</td>
<td></td>
</tr>
<tr>
<td>C04-017</td>
<td>He</td>
<td></td>
<td>56.3</td>
<td>2.0</td>
<td></td>
</tr>
<tr>
<td>C04-018</td>
<td>He</td>
<td></td>
<td>48.3</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td>C04-020</td>
<td>He</td>
<td></td>
<td>63.3</td>
<td>1.9</td>
<td></td>
</tr>
</tbody>
</table>

* Slip rate estimate from ALR fan data is based on the reconstructed displacement of 1110±110 m and the age of an erosion resistant quartz sample Calico-9 (346±24 ka). Slip rate estimate from the TR alluvial fan is based on the displacement estimate of 90 m and weighted mean age of 14.0±5.8 ka (using two standard deviations as uncertainty) from TCN dating of four rock samples with apparent ages between 10.7 ka and 17.1 ka. Using OSL dating ages result in higher rates. Slip rate estimate by Selander (2015) is based on a depth profile Q2c at the southwest Rodman Mountains (SWRM), with a displacement estimate of 24±10 m and an age estimate of 17.1±5.2 ka. Slip rate estimate of B fan by Oskin et al. (2007a, b) is from an assigned age of 650±100 ka based on age data (see details in Oskin et al. (2007a)) and a displacement estimate of 900±200 m. Slip rate estimate of K fan by Oskin et al. (2007a) is from a displacement estimate of 110±10 m and a mean average age of 56.4±4.7 ka (inheritance estimate from active wash samples subtracted) based on 10Be exposure age dating of rock samples.
Figure 1. Fault map showing the ECSZ in the Mojave Desert, from U.S. Geological Survey and California Geological Survey (2006). Color indicates time of recent movement. Blue triangles show locations of geologic strike-slip rate estimates from recent studies, with rates given in mm/yr (Oskin and Iriondo, 2004; Oskin et al., 2007a, 2008; Selander, 2015). Beach balls mark the 1992 Mw 7.3 Landers earthquake and the 1999 Mw 7.1 Hector Mine earthquake. Blue square shows our study site. Dashed magenta box outlines the area of Figure 15b. Selected fault names are labeled beside corresponding faults. Insert: Grey outlines area shown in the main figure. Red dot marks Los Angeles. Magenta square indicates location of the outlined magenta box in the main figure (corresponding to the area shown in Figure 15b).
Figure 2. Summed geodetic and geologic rates across the ECSZ. Black line and gray area mark the summed geologic slip rate and uncertainty (95% confidence) from Oskin et al. (2008). Blue dots with error bars represent summed geodetic rates and their uncertainties (also 95% confidence) from elastic deformation models (a) – (k). Slip rates for (d) (McCaffrey, 2005) and (h) (Loveless and Meade, 2011) calculated at latitude 34.8° N based on their block models. Spiniér et al. (2010) and Liu et al. (2015) provide multiple solutions. Miller et al. (2001) and Evans et al. (2016) did not provide uncertainties.
Figure 3. Images of the study area. (a) Aerial ortho-image. (b) LiDAR hillshade overlain on aerial image. Blue boxes outline two offset landforms shown in Figure 5 (ALR alluvial fan) and 7 (TR alluvial fan). Dashed white lines indicate mapped fault traces, dashed black lines are inferred fault traces. Red dots show locations of rock samples. Light blue squares are trench or locations: CalicoA and Calico-Pit3 on the ALR alluvial fan and Calico-Pit2 on the TR alluvial fan. Coordinates are in UTM zone 11 N.
Figure 4. Slip restoration for the ALR alluvial fan. (a) Aerial image of the ALR alluvial fan. (b) The pair of red and yellow dots marked key features used to align offset fan bodies, width of the major channel is used as uncertainty. (c) Restored alluvial fan. (d) Hillshade of the restored alluvial fan.
Figure 5. Displacement of the ALR alluvial fan (offset 1 outlined in Figure 3b). (a) Aerial image of the ALR alluvial fan, red dots mark prominent drainages on alluvial fan surfaces, yellow and cyan dots mark width of major channel used to define uncertainty of displacement estimate. (b) LiDAR hillshade overlain on aerial image, note the prominent drainage on offset fan bodies. (c) Slope aspect derived from LiDAR DEM overlain on aerial image. (d-f) Restoration of 1110 m of right-lateral slip on the Calico Fault, corresponding to present surface features shown in (a-c). (g-h) Elevation contour (white lines) before and after restoration overlain on the hillshade, contour interval is 2 m.
Figure 6. Displacement restoration using LaDiCaOZ software (Zielke et al., 2015; Haddon et al., 2016) for the ALR alluvial fan. (a) Contour of the ALR alluvial fan, 5 m contour interval. Cyan line marks the fault trace, blue line marks an elevation profile shown in (b), red line marks another profile with elevation between the two yellows dots shown in (a). Red and blue dotted lines mark the upstream and downstream limits of the prominent drainage in the fan bodies. (b) Blue line corresponds to the blue line in (a); red line corresponds to the section of the red profile marked by two yellow dots in (a). (c) Best match by shifting the red profile in (b); (d–e) shows misfit using different displacements. The best match corresponds to a horizontal displacement of 1111 m. (f) Reconstructed elevation map based on a 1111 m displacement, contours at two ends have some artificial distortion due to the algorithm in the software but will not affect the analysis. (g) Red and blue scatter corresponding to dotted red and blue lines in (a); magenta lines are visually fits of straight lines. Black vertical line corresponds to fault location. Dashed red and blue lines correspond to the red and blue line in (a). (h) Profile of the dominant drainage after restoration. Black line is a reference to examine the straightness of the reconstructed dominant drainage. (i–j) Red and blue scatter corresponding to dashed red and blue lines in (a), the other symbols are the same as in (g) and (h).
Figure 7. Displacement of the TR alluvial fan (offset 2 outlined in Figure 3b). (a) Aerial image of the TR alluvial fan. (b) LiDAR hillshade overlain on aerial image. (c) Slope aspect derived from LiDAR DEM overlain on aerial image. (d) Elevation contour (white lines) overlain on the hillshade, contour interval is 2 m. (e-h) Restoration of 20 m displacement by aligning the TR-ASC boundary with an active channel on the northeastern wall of the Calico Fault (see dashed cyan line in Figure 8a). This assumes the boundary of the TR alluvial fan was originally highly nonlinear (Figures 8a, 8b). (i-l) Restoration of 90 m displacement when aligning linear downstream channel with oldest upstream fan edge, assuming the ASC incised completely into the IAF surface (Figure 8g). (m-p) Restoration of 200 m displacement assuming the anastomosing stream channel (ASC) surface incised completely into the TR alluvial fan (Figure 8h).
Figure 8. Different restorations of the offset at the TR alluvial fan. (a-d) Zoomed in images outlined by red boxes in Figures 7 a, e, i, m. Red line in (a) marks the 20 m displacement for the restoration in (b). (e) An enlarged slope aspect map of Figure 7c, trends of paleostream channels within the dashed shape (IAF*) are the same as the broader intermediate-aged alluvial fan surface (IAF). (f) Two elevation profiles marked by straight golden and black lines in (e). (g, h) show two different restorations with different units shaded.
Figure 9. Displacement restoration using LaDiCaoz (Zielke et al., 2015; Haddon et al., 2016) for the TR alluvial fan. Markers in each subplot correspond to the same parameters described in Figure 6. Note that due to the complexity of the fan surfaces, we do not pick points near the fault to define the offset profiles, instead, we use the linear traces far from the fault to define an offset channel and extend them onto the fault. Displacement restoration from this fan is 169 m. Note that the reconstructed channel in (h) cannot be well fit by a straight line, probably due to uneven subsidence or erosion, or incorrect restoration since desert flow seems to have significantly modified the surface feature.
Figure 10. Age constraints from stony rubble packing on the alluvial fan surface and carbonate rind thickness. (a) Percentages of pebble coverage for flow surface with different ages, reflecting degree of packing on the desert pavement. Black dots with error bar show data depicted in Wells et al. (1985), pink corresponds to the range of pebble coverage percentage for the ALR alluvial fan surface (86%-97%, calculated using photos taken during our field investigation, see Supplementary Figures S3-S6). Dashed red line marks the corresponding age of the intercept between a piecewise model and the lower bound of pebble coverage. (b) Histogram of carbonate rind thickness for samples collected from evacuated pits and exposures rocks at a scarp. Thicknesses are measured under a ×10 binocular scope, using a stainless steel ruler with 1 mm scale (see example in Supplementary Figure S2). Dashed red lines mark the mode/median/max thickness, annotations are corresponding ages calculated using the model of Amoroso (2006). Solid red line marks calculated thickness based on the preferred age for ALR alluvial fan from 10Be exposure dating, pink area marks the uncertainty.
Figure 11. TCN $^{10}$Be dated ages (ka; see Table 2 for the age uncertainties) of rock samples collected on alluvial fan surfaces. (a) Red dots show locations of rock samples. Their apparent ages are shown in adjacent white annotations. (b) Age probability density function (PDF) of rock samples on the ALR alluvial fan, derived by using the program of Zechar and Frankel (2009). (c) Age PDF of rock samples on the TR alluvial fan.
Figure 12. $^{10}$Be versus depth, and corresponding Bayesian-Monte Carlo simulations. (a) through (e) show results for profile CalicoA. In (a), blue markers with error bar show measured $^{10}$Be concentration of samples, grey curves are from 100,000 Bayesian-Monte Carlo simulations following Hidy et al. (2010) and Hidy (2013), red curve shows best fit. (b) shows age versus erosion rate, each red dot represents one possible solution, the scatter defines the solution space. Blue dots mark the top 100 fits (lowest chi-square). Red lines in (c-e) show the age probability density functions (PDF) of age, erosion rate and inheritance from the simulation, dashed black lines shows relative value of the minimum chi-square. Note that one sample collected at 50 cm depth for CalicoA (marked by blue triangle in (a)) was not used because its concentration lies well off the exponential trend. Inclusion of this sample yields a higher erosion rate and inheritance from the simulation. (c) shows the age probability density functions (PDF) of age, erosion rate and inheritance from the simulation, dashed black lines shows relative value of the minimum chi-square. Note that one sample collected at 50 cm depth for CalicoA (marked by blue triangle in (a)) was not used because its concentration lies well off the exponential trend. Inclusion of this sample yields a higher erosion rate and inheritance from the simulation. (d) shows the age probability density functions (PDF) of age, erosion rate and inheritance from the simulation, dashed black lines shows relative value of the minimum chi-square. Note that one sample collected at 50 cm depth for CalicoA (marked by blue triangle in (a)) was not used because its concentration lies well off the exponential trend. Inclusion of this sample yields a higher erosion rate and inheritance from the simulation. (e) shows the age probability density functions (PDF) of age, erosion rate and inheritance from the simulation, dashed black lines shows relative value of the minimum chi-square. Note that one sample collected at 50 cm depth for CalicoA (marked by blue triangle in (a)) was not used because its concentration lies well off the exponential trend. Inclusion of this sample yields a higher erosion rate and inheritance from the simulation. (f-j) Corresponding Bayesian-Monte Carlo simulations for profile Calico-Pit3. Sample at 30 cm depth (blue triangle in (f)) was excluded from the simulations as it lies well off the expected exponential trend. Because a conservative criterion for this depth profile would yield unrealistically high end-members of inheritance, we used the maximum $^{10}$Be concentration among the 6 samples to define the upper limit of the inheritance for the simulations. (k) Sample depth versus $^{10}$Be concentration for Calico-Pit2 trench in the TR alluvial fan. Lack of exponential decrease with depth suggests that this fan has undergone complex deposition-erosion history. Note for CalicoA, error bars in (a) show 5σ confidence level. This was necessary to run the simulator due to the data scatter (see also Hedrick et al. (2017)). Error bars in (f) and (k) show 2σ confidence level.
Figure 13. Age constraints for the ALR and TR alluvial fan. (a) $^{10}$Be dated age of 345±24 ka is our preferred age estimate for the ALR alluvial fan. Degree of desert pavement packing, carbonate rind thickness and soil chronostratigraphy give additional constraints on the fan age. (b) The youngest OSL dated age is a minimum age for the TR alluvial fan, and the $^{10}$Be dated contains an unknown amount of inheritance.

Figure 14. Plausible displacement for the TR alluvial fan offset. Dashed blue lines mark the minimum and maximum possible ages from OSL and TCN $^{10}$Be dating. Red lines show the minimum and maximum possible slip rates for the Calico Fault (Sauber et al., 1994; Oskin et al., 2007a; Selander, 2015; McGill et al., 2015). Grey area represent the range of plausible displacement for the TR alluvial fan offset, between 7–168 m. Black dots with error bars mark different restorations in this study.
Figure 15. Along-strike slip rate estimates on the Calico Fault. (a) Color represents age. ALR: 3.2±0.4 mm/yr from this study; B: 1.4±0.3 mm/yr from the ‘B’ alluvial fan in Oskin et al. (2007a); K: 1.8±0.3 mm/yr from the ‘K’ alluvial fan in Oskin et al. (2007a); Q2c: 1.4±0.3 mm/yr from the ‘Q2c’ depth profile in Selander (2015). (b) Fault map shows the area outlined by dashed magenta box in Figure 1. Blue circles mark the geologic sites corresponding to (a). Light blue circle shows location of the TR alluvial fan. Black lines show major fault traces from U.S. Geological Survey and California Geological Survey (2006).
Figure 16. Ages of dated samples from Oskin et al. (2007a, b), Selander (2015) and this study, grouped in older (B and ALR) and younger (K, Q2c, TR) alluvial fans. Black dots with error bars show $^{10}$Be dated ages, triangles with error bars show $^3$He dated ages, squares with error bars show $^{40}$Ar/$^{39}$Ar dated ages, note that some age error bars are smaller than the markers. Red squares with error bars show preferred estimates, note that Q2c unit age is from a depth profile. Note that ‘B’ fan studied by Oskin et al. (2007a, b) uses data from three techniques: $^{40}$Ar/$^{39}$Ar, $^3$He and $^{10}$Be. Outliers near the beginning and end of $^{40}$Ar/$^{39}$Ar step-heating runs are omitted. Also note that the 90 m displacement for the TR alluvial fan is a estimate by aligning the northwestern edges of the TR alluvial fan surfaces (Figures 7i-7l, 8c, 8g). Dashed blue line marks the TR alluvial fan age from OSL dating (5 ka). Locations of these fans are shown in Figure 15b. Detail of these ages are in Table 2.

Figure 17. Hypothetical apparent age distribution for alluvial fans of different ages based on surface exposure dating using the TCN method. (a) Probability of sample ages from a relatively young fan, black curve shows a chi-square distribution reflecting increased likelihood of inheritance. (b) Probability of sample ages from a relatively old fan, black curve shows a chi-square distribution, reflecting increasing likelihood of surface disturbance. Note the scale of age in (b) is different from (a).
Figure 18. Age and displacement estimates for the Calico Fault from Oskin et al. (2007a, b), Selander (2015) and this study. (a) Black dots with error bars show \( ^{10}\)Be dated ages, triangles with error bars show \( ^3\)He dated ages, squares with error bars show \( ^{40}\)Ar/\( ^{39}\)Ar dated ages. Red squares with error bars show preferred estimates. Note that some age error bars are smaller than the markers. Insert box expands data near origin. (b) Preferred estimates only, same as red square markers shown in (a). Grey lines show different mean slip rates for reference.

Figure 19. Slip rate models based on total displacement (9.8 km) and initiation age of the Calico Fault. (a) Slip rate estimates for a fault initiation age of 4.1 Ma. Grey line represents a constant slip rate of 1.8 mm/yr, the total displacement is 7.4 km. Each black line shows slip rate increasing with a constant acceleration speed: for present-day slip rates of 1.8 mm/yr and 3.2 mm/yr, the total displacements are 3.7 km and 6.6 km, respectively; to have a total displacement of 9.8 km, this model requires a present-day slip rate equals to 4.8 mm/yr. (b) Slip rate variation for an initiation age 4.1 Ma with different Rayleigh scale parameter \( S \) (time between fault initiation and maximum acceleration) based on a damage growth model (Gourmelen et al., 2011). \( R_g \) represents present-day slip rate. (c) Present-day slip rate based on plausible fault initiation age \( t_0 \) and time between fault initiation and maximum acceleration in speed \( S \). Numbers (in mm/yr) mark selected slip rate contour lines.
Figure 20. Comparison of geodetic slip rate and scaled geologic slip rate estimates. (a) Black dots with error bars (95% confidence) show on-fault geologic slip rate estimates of the six major active dextral faults (from west to east) from Oskin et al. (2007a, 2008). Red dot with error bar (also 95% confidence) shows our preferred slip rate estimate based on the ALR alluvial fan data. Yellow dots show scaled slip rates by assuming off-fault deformation accounts for 40±23% (Herbert et al., 2014a) of the total slip and have the same percentages for all individual faults, error bars omitted for clarity. (b) Blue dots with error bars show summed geodetic slip rate estimates, corresponding to references shown in Figure 2. Black line and gray area mark the summed geologic slip rate and uncertainty (95% confidence) from Oskin et al. (2008). Red line and pink area mark the updated summed geologic slip rate and uncertainty (95% confidence) based on our new slip rate estimated from the ALR alluvial fan, plus scaled slip rate estimates from Oskin et al. (2008), assuming off-fault deformation accounts for 40±23% (Herbert et al., 2014a) of total slip. The new accumulative geologic rate across the ECSZ overlaps with most of the geodetic rate estimates within uncertainty.
Appendix V: Copyright permission for Xie et al., 2019, JGR

Permissions policy for publications | AGU

PERMISSIONS POLICY

AGU grants permission for individuals to use figures, tables, and short quotes from AGU journal and books for republication in academic works and to make single copies for personal use in research, study, or teaching provided full attribution is included. There is no need to request this permission from AGU.

This permission does not extend to public posting of the PDF or HTML created by AGU for publication. AGU journal content after 1996 is now freely available 24 months after publication; AGU encourages linking to this content directly. There is no charge or additional permissions needed for any of these uses, but the material must be cited appropriately.

For information on requesting permission for commercial reuse of AGU content, please click on the "permissions" link on any journal or book home page in the Wiley Online Library (https://agupubs.onlinelibrary.wiley.com/).

Rights granted to authors

AGU’s philosophy recognizes the need to ensure that authors have a say in how their works are used and the necessity to foster broad dissemination of scientific literature while protecting the viability of the publication system. The following nonexclusive rights are granted to AGU authors:

https://www.agu.org/Publish-with-AGU/Publish/Author-Resources/Policies/Permission-policy
• All proprietary rights other than copyright (such as patent rights)
• The right to present the material orally
• The right to reproduce figures, tables, and extracts properly cited
• The right to make paper copies of all or part of the contribution for classroom use
• The right to deny subsequent commercial use of the contribution
• The right to place the contribution or its abstract on his/her personal website as described below.

Policies

Permission to deposit an article in an institutional repository

*Adopted by Council 13 December 2009*

AGU allows authors to deposit their journal articles if the version is the final published citable version of record, the AGU copyright statement is clearly visible on the posting, and the posting is made 6 months after official publication by the AGU.

Permission to post an article to author's website

Authors may post their unformatted papers or their abstracts to their own Web sites or their departmental Web sites according to the guidelines listed below. If authors wish to post preprints of their articles on other sites, they should be aware of the relevant part of the *Dual publication policy* (/Publish-with-AGU/Publish/Author-Resources/Policies/Prior-Publication-Policy) that deals with preprints:

*Dual Publication Policy:* "AGU does allow posting of preprints and accepted papers in not-for-profit preprint servers that are designed to facilitate community engagement and discovery across the sciences."

1. If the paper has been submitted for publication, but not yet accepted, the author should include the following statement if he/she places the paper on a Web site:
   • "Submitted for publication in (journal title)."

2. If the paper has been accepted for publication and copyright has been transferred to AGU, the author may place the paper on his/her own Web site with the following statement appearing on the first screen of the abstract or article:
   • "Accepted for publication in (journal title). Copyright (year) American Geophysical Union. Further reproduction or electronic distribution is not permitted."

3. If the paper has been published, or when it is published, the above statement should be changed to the following:
   • "An edited version of this paper was published by AGU. Copyright (year) American Geophysical Union."

It is recommended that the full citation and a link to the open abstract also be provided:
“Author(s), Year of publication (in parentheses), Title of article, Name of journal, Volume number, Citation number, Digital Object Identifier (DOI). To view the published open abstract, go to http://dx.doi.org (http://dx.doi.org/) and enter the DOI.”

Providing the full bibliographic reference helps assure that authors receive full credit through correct citation to their articles, while including the URL directs viewers to the published abstract.

4. If an article was placed in the public domain, the following statement should appear on the first screen of the abstract or article:

“Not subject to U.S. copyright”. Please substitute “published” for the word “copyright” in the credit line mentioned in items 2 and 3.
Appendix VI: Error propogations

Considering errors in GPS-buoy geometry used for anchor position estimates, and substituting the three rotation matrix, Equation 2.2 becomes:

\[
\begin{bmatrix}
N_{ag} \\
E_{ag} \\
U_{ag}
\end{bmatrix} =
\begin{bmatrix}
\cos(\alpha) \cos(\beta) & \cos(\alpha) \sin(\beta) \sin(\gamma) - \sin(\alpha) \cos(\gamma) & \cos(\alpha) \sin(\beta) \cos(\gamma) + \sin(\alpha) \sin(\gamma) \\
\sin(\alpha) \cos(\beta) & \sin(\alpha) \sin(\beta) \sin(\gamma) + \cos(\alpha) \cos(\gamma) & \sin(\alpha) \sin(\beta) \cos(\gamma) - \cos(\alpha) \sin(\gamma) \\
-\sin(\beta) & \cos(\beta) \sin(\gamma) & \cos(\beta) \cos(\gamma)
\end{bmatrix}\begin{bmatrix}
p \\
q \\
L
\end{bmatrix}
\] (A1)

Assuming \(p, q, L, \alpha, \beta, \) and \(\gamma\) are independent parameters, partial derivatives of Equation A1 are shown in Equations A2–A4:

\[
\begin{align*}
\frac{\partial N_{ag}}{\partial p} &= \cos(\alpha) \cos(\beta) \\
\frac{\partial N_{ag}}{\partial q} &= \cos(\alpha) \sin(\beta) \sin(\gamma) - \sin(\alpha) \cos(\gamma) \\
\frac{\partial N_{ag}}{\partial L} &= \cos(\alpha) \sin(\beta) \cos(\gamma) + \sin(\alpha) \sin(\gamma) \\
\frac{\partial N_{ag}}{\partial \alpha} &= -p \sin(\alpha) \cos(\beta) - q \sin(\alpha) \sin(\beta) \sin(\gamma) - q \cos(\alpha) \cos(\gamma) - L \sin(\alpha) \sin(\beta) \cos(\gamma) \\
&\quad + L \cos(\alpha) \sin(\gamma) \\
\frac{\partial N_{ag}}{\partial \beta} &= -p \cos(\alpha) \sin(\beta) + q \cos(\alpha) \cos(\beta) \sin(\gamma) + L \cos(\alpha) \cos(\beta) \cos(\gamma) \\
\frac{\partial N_{ag}}{\partial \gamma} &= q \cos(\alpha) \sin(\beta) \sin(\gamma) + q \sin(\alpha) \sin(\gamma) - L \cos(\alpha) \sin(\beta) \sin(\gamma) + L \sin(\alpha) \cos(\gamma)
\end{align*}
\] (A2)
\[
\begin{align*}
\frac{\partial E_{ak}}{\partial p} &= \sin(\alpha)\cos(\beta) \\
\frac{\partial E_{ak}}{\partial q} &= \sin(\alpha)\sin(\beta)\sin(\gamma) + \cos(\alpha)\cos(\gamma) \\
\frac{\partial E_{ak}}{\partial L} &= \sin(\alpha)\sin(\beta)\cos(\gamma) - \cos(\alpha)\sin(\gamma) \\
\frac{\partial E_{ak}}{\partial \alpha} &= p\cos(\alpha)\cos(\beta) + q\cos(\alpha)\sin(\beta)\sin(\gamma) - q\sin(\alpha)\cos(\gamma) + L\cos(\alpha)\sin(\beta)\cos(\gamma) \\
&\quad + L\sin(\alpha)\sin(\gamma) \\
\frac{\partial E_{ak}}{\partial \beta} &= -p\sin(\alpha)\sin(\beta) + q\sin(\alpha)\cos(\beta)\sin(\gamma) + L\sin(\alpha)\cos(\beta)\cos(\gamma) \\
\frac{\partial E_{ak}}{\partial \gamma} &= q\sin(\alpha)\sin(\beta)\cos(\gamma) - q\cos(\alpha)\sin(\gamma) - L\sin(\alpha)\sin(\beta)\sin(\gamma) - L\cos(\alpha)\cos(\gamma) \\
\end{align*}
\] (A3)

\[
\begin{align*}
\frac{\partial U_{ak}}{\partial p} &= -\sin(\beta) \\
\frac{\partial U_{ak}}{\partial q} &= \cos(\beta)\sin(\gamma) \\
\frac{\partial U_{ak}}{\partial L} &= \cos(\beta)\sin(\gamma) \\
\frac{\partial U_{ak}}{\partial \alpha} &= 0 \\
\frac{\partial U_{ak}}{\partial \beta} &= -p\cos(\beta) - q\sin(\beta)\sin(\gamma) - L\sin(\beta)\cos(\gamma) \\
\frac{\partial U_{ak}}{\partial \gamma} &= q\cos(\alpha)\cos(\gamma) - L\cos(\beta)\sin(\gamma) \\
\end{align*}
\] (A4)
After ignoring items that are insignificant in total error budgets, formulas A5–A7 show the maximum range of possible errors induced by corresponding parameters. In these formulas, $\Delta N_{ag}|_{\Delta p}$ denotes error in northern component of the anchor position caused by error in $p$, and so forth. Note that all items are simplified to be positive values in formulas A5–A7, pitch and roll measurements between $-2^\circ$ and $2^\circ$ are used. The derived values are maximum possible errors.

\[
\begin{align*}
\Delta N_{ag}|_{\Delta p} & \leq \Delta p \\
\Delta N_{ag}|_{\Delta q} & \leq [\sin(\alpha)] \Delta q \leq \Delta q \\
\Delta N_{ag}|_{\Delta L} & \leq [\sin(\beta) + \sin(\gamma)] \Delta L < [\sin(2^\circ) + \sin(2^\circ)] \Delta L = 0.07 \Delta L \\
\Delta N_{ag}|_{\Delta \alpha} & \leq [p + q + L \sin(\beta) + L \sin(\gamma)] \Delta \alpha < [p + q + L \sin(2^\circ) + L \sin(2^\circ)] \Delta \alpha \\
& = [p + q + 0.07L] \Delta \alpha \\
\Delta N_{ag}|_{\Delta \beta} & \leq [p \sin(\beta) + q \sin(\gamma) + L] \Delta \beta < [p \sin(2^\circ) + q \sin(2^\circ) + L] \Delta \beta \\
& = [0.035p + 0.035q + L] \Delta \beta \\
\Delta N_{ag}|_{\Delta \gamma} & \leq [q \sin(\beta) + q \sin(\gamma) + L] \Delta \gamma < [q \sin(2^\circ) + q \sin(2^\circ) + L] \Delta \gamma = [0.07q + L] \Delta \gamma
\end{align*}
\]
\[
\begin{align*}
\Delta E_{ag}|_{\Delta p} & \leq [\sin(\alpha)]\Delta p \leq \Delta p \\
\Delta E_{ag}|_{\Delta q} & \leq \Delta q \\
\Delta E_{ag}|_{\Delta L} & \leq [\sin(\beta) + \sin(\gamma)]\Delta L < [\sin(2^\circ) + \sin(2^\circ)]\Delta L = 0.07\Delta L \\
\Delta E_{ag}|_{\Delta \alpha} & \leq [p + q + L\sin(\beta) + L\sin(\gamma)]\Delta \alpha < [p + q + L\sin(2^\circ) + L\sin(2^\circ)]\Delta \alpha \\
& = [p + q + 0.07L]\Delta \alpha \\
\Delta E_{ag}|_{\Delta \beta} & \leq [p\sin(\beta) + q\sin(\gamma) + L]\Delta \beta < [p\sin(2^\circ) + q\sin(2^\circ) + L]\Delta \beta \\
& = [0.035p + 0.035q + L]\Delta \beta \\
\Delta E_{ag}|_{\Delta \gamma} & \leq [q\sin(\beta) + q\sin(\gamma) + L]\Delta \gamma < [q\sin(2^\circ) + q\sin(2^\circ) + L]\Delta \gamma = [0.07q + L]\Delta \gamma \\
\end{align*}
\] (A6)

\[
\begin{align*}
\Delta U_{ag}|_{\Delta p} & < [\sin(2^\circ)]\Delta p = 0.035\Delta p \\
\Delta U_{ag}|_{\Delta q} & \leq [\sin(\gamma)]\Delta q < [\sin(2^\circ)]\Delta q = 0.035\Delta q \\
0.9988\Delta L & = [\cos(2^\circ)\cos(2^\circ)]\Delta L < \Delta U_{ag}|_{\Delta \alpha} \leq \Delta L \\
\Delta U_{ag}|_{\Delta \alpha} & = 0 \\
\Delta U_{ag}|_{\Delta \beta} & \leq [p + L\sin(\beta)]\Delta \beta < [p + L\sin(2^\circ)]\Delta \beta = [p + 0.035L]\Delta \beta \\
\Delta U_{ag}|_{\Delta \gamma} & \leq [q + L\sin(\gamma)]\Delta \gamma < [q + L\sin(2^\circ)]\Delta \gamma = [q + 0.035L]\Delta \gamma \\
\end{align*}
\] (A7)
Given input errors as: $p = \pm 3 \text{ cm}$, $q = \pm 3 \text{ cm}$, $\Delta L = \pm 30 \text{ cm}$, $\Delta \alpha = \pm 0.3^\circ$, $\Delta \beta = \pm 0.2^\circ$, $\Delta \gamma = \pm 0.2^\circ$. Formulas A8–A10 show the range of propagated errors in anchor position estimates.

\[
\begin{align*}
\Delta N_{ag}\mid \Delta p & \leq 0.030 \text{ m} \\
\Delta N_{ag}\mid \Delta q & \leq 0.030 \text{ m} \\
\Delta N_{ag}\mid \Delta L & < 0.021 \text{ m} \\
\Delta N_{ag}\mid \Delta \alpha & < 0.010 \text{ m} \\
\Delta N_{ag}\mid \Delta \beta & < 0.096 \text{ m} \\
\Delta N_{ag}\mid \Delta \gamma & < 0.096 \text{ m}
\end{align*}
\tag{A8}
\]

\[
\begin{align*}
\Delta E_{ag}\mid \Delta p & \leq 0.030 \text{ m} \\
\Delta E_{ag}\mid \Delta q & \leq 0.030 \text{ m} \\
\Delta E_{ag}\mid \Delta L & < 0.021 \text{ m} \\
\Delta E_{ag}\mid \Delta \alpha & < 0.010 \text{ m} \\
\Delta E_{ag}\mid \Delta \beta & < 0.096 \text{ m} \\
\Delta E_{ag}\mid \Delta \gamma & < 0.096 \text{ m}
\end{align*}
\tag{A9}
\]

\[
\begin{align*}
\Delta U_{ag}\mid \Delta p & < 0.001 \text{ m} \\
\Delta U_{ag}\mid \Delta q & < 0.001 \text{ m} \\
0.2996 \text{ m} & < \Delta U_{ag}\mid \Delta L < 0.300 \text{ m} \\
\Delta U_{ag}\mid \Delta \alpha & = 0 \text{ m} \\
\Delta U_{ag}\mid \Delta \beta & < 0.003 \text{ m} \\
\Delta U_{ag}\mid \Delta \gamma & < 0.003 \text{ m}
\end{align*}
\tag{A10}
\]
Figure A1: Inverse perspective transformation of photographic images taken by a UAV on 23 August 2018 (GPS-buoy deployment day). Upper right annotations on the right column are estimated compass heading directions.
Figure A2: Inverse perspective transformation of photographic images taken on 9 September 2018. (a1, b1) are photos taken by a cellphone. (a2, b2) are projected maps. All color and markers denote the same meanings as Figure A1, except that the azimuth directions are both parallel to the forward direction of the compass’ heading. Note that the GPS antenna and meteorological sensor (marked by two pelicans) form a line that is nearly parallel to the shoreline of the Egmont Key, different than Figure A1, indicating a change in digital compass heading direction.
Figure A3: Inverse perspective transformation of photographic images taken on 5 April 2019. All color and markers denote the same meanings as Figure A1.
Figure A4: Anchor displacement and environmental variables during periods of relatively large motion in May and June 2019. Note that light blue shade marks the time of large displacement during falling tide when current speed is at local maximum.
Figure A5: GPS position time series of a land site for comparison. ZEFR (operated by CORS network: https://www.ngs.noaa.gov/CORS/) is a land site used for comparison, its kinematic position time series (15 second interval) are shown in grey on the North/East/Up panels. Data processing following the same procedure as for the GPS-buoy site described in the main paper. A large data gap in late March 2019 is due to lack of observations at ZEFR station. Orange dots are detected outliers using the modified Z-score method describe in the main paper. Red dots with error bars show daily medians and uncertainties using the same method described in chapter 2. Weighted one standard derivations (SD) of different components during a ∼5 months period (same as the period shown in Figure 2.12) are annotated in red.
Figure A6: Illustration of error in vertical component of anchor position estimates induced by error in GPS to anchor length measurement. \( L \) represents the actual buoy length, \( L' \) represents buoy length used in anchor position estimates. Error in buoy length (\( \Delta L \)) causes an error in anchor vertical position estimate (\( \Delta U \)) that is systematic and will not significantly affect displacement estimates.

Figure A7: Water temperature measured near the surface. Data collected at the water level station shown by orange square in 2.1b.
Figure A8: Anchor displacement and current velocities along the long-term motion direction during Hurricane Michael (a) and the unnamed storm in December 2018 (b). Anchor long-term motion direction is estimated by fitting a straight line to daily time series during a period 23 August 2018 to 1 May 2019 (colored dots on the right column of Figure 13 in the main paper). Second row shows observed (location shown in Figure 2.1b) and modeled current velocities projected along anchor long-term motion direction. Third row shows current velocities projected along mean local surface outgoing flow direction, estimated by fitting a straight line to GPS trajectories during the corresponding period.
Figure A9: Simulation of effect of extreme weather condition at seafloor ballast. (a) Static seabed situation. The ballast cannot move due to large residual friction (FF). (b) Assuming a weak layer between the ballast and the seafloor. The ballast can move during extreme weather conditions in (b), but not in (a). A seafloor slope of 3.5° is used in both (a) and (b).