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Abstract

Surface velocities and strain rates from satellite geodesy have become essential tools for understanding the distribution of tectonic deformation, faulting and seismic hazard. However, across large regions of distributed continental deformation, such as the Alpine-Himalayan Belt, data are only sparsely available. While previous studies have mainly used spatially sparse GNSS to measure deformation at such large scales, these approaches cannot characterize shorter wavelength features of deformation in many places. We use Sentinel-1 radar images acquired during 2016-2024 to provide trans-national average surface velocities and time series at 1 km spatial resolution stretching a distance of over 11,000 km from southern Europe to eastern China, covering an area more than 20 million square kilometres. We produce the velocity field by combining data from over 220,000 Sentinel-1 SAR images with a new belt-wide compilation of GNSS velocities, all combined in a consistent Eurasian reference frame. Horizontal strain rates are derived from gradients of the velocity field, yielding near-continuous spatial deformation information over the entirety of the largest deforming region on the planet. The horizontal velocities and strains are dominated by tectonic deformation, which has a bimodal behaviour – focused on major faults but distributed elsewhere. Shorter-wavelength vertical velocities are dominated by non-tectonic processes, in par-

ticular the widespread over-exploitation of groundwater. Our new velocity and strain rates are foundational data sets that reveal the details of how the continents deform for the first time at trans-continental scale.
Keywords: Radar remote sensing, GNSS, Sentinel-1, InSAR, Active Tectonics, Continental Deformation, Faulting, Strain, Seismic Hazard, Earthquakes

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1. Introduction

The collision of the African, Arabian, and Indian plates with Eurasia has created the Alpine Himalayan Belt (AHB), a vast region of thickened crust and mountain ranges spanning the southern margin of the Eurasian continent (Figure 1) and reaching ∼2,000 km into its interior in places. This major orogeny creates, supports and shapes the highest mountains in the world and has been a long-standing focus of research on the deformation of the continents (Molnar and Tapponnier, 1978; England and Jackson, 1989). The region is marred by numerous active seismogenic faults (Styron and Pagani, 2020), generating some of the largest, and often most fatal, earthquakes (Bilham, 2019) – 75% of earthquakes that have killed more than 10,000 people since 1900 have occurred in the AHB (England and Jackson, 2011). Understanding the kinematics of this range is important for constraining the style of tectonics of the continents (Jackson and McKenzie, 1984; Yin, 2010), which in turn act as first-order constraints on dynamical models of deformation (England and Molnar, 1997; Vergnolle et al., 2007) and how the continents and margins might evolve through time (McKenzie, 2025). Furthermore, deformation data are an increasingly important input for seismic hazard assessments, providing important independent constraints on the slip rates of major faults and spatial maps of moment accumulation rates (Bird et al., 2015; Field et al., 2015; Stevens and Avouac, 2021).

The recent explosion in the availability of remotely-sensed data from satellite platforms, in particular in the past decade by the European Space Agency's Sentinel-1 radar mission (Salvi et al., 2012; Potin et al., 2016), has enabled a step change in the scale of Interferometric Synthetic Aperture Radar (InSAR) measurements (Gabriel et al., 1989). The open Sentinel-1 archive and advances in big data processing (e.g., Lazecký et al., 2020) have enabled us to produce the first transcontinental, decadal time series of surface velocities at a resolution of 1 km. This allows us to resolve the details of tectonic processes at scale. From the Sentinel-1 data, and new compilation of Global Navigation Satellite System (GNSS) data, we derive average three-component surface velocities in a consistent Eurasian reference frame, and horizontal surface strain rates. The results show the balance in tectonic deformation between regions of high strain localisation and those with more distributed deformation, as well as highlighting non-tectonic processes, particularly those associated with exploitation of groundwater. We make these new time series, velocities and strain rates freely available for further analysis and modelling by the wider community.

1.1. Motivation: Understanding the deformation of the continents

Understanding the present-day deformation of the continents is important for assessing the distribution of seismic hazard through the calculation of strain rates (Ward, 1998; Bird et al., 2010), and for constraining the kinematics of crustal tectonics to allow testing and assessment of the potential dynamics of continental deformation, (e.g., Flesch et al., 2001; Wang and Barbot, 2023; Fang et al., 2024a).

It has long been recognized that plate tectonics operates differently in the continents than in the oceans (McKenzie, 1972; Molnar and Tapponnier, 1975; McKenzie, 2025), but understanding how continents deform has been hampered by the lack of observations with the spatial resolution required to resolve tectonic processes. Understanding where and why contrasting styles of deformation are observed remain key questions in active tectonics (Thatcher, 2007). In particular, we want to understand why seismicity and strain in some regions appear largely focused on large block-bounding faults such as the North Anatolian Fault (Weiss et al., 2020), similar to the behaviour seen in oceanic lithosphere, whereas other regions appear to have more diffuse strain and seismicity (Watson et al., 2024). This is useful for calculating the distribution of hazard (Liu and Stein, 2016), and determining the degree to which continents experience on-fault versus off-fault deformation (Zeng and Shen, 2016). We want to be able to determine

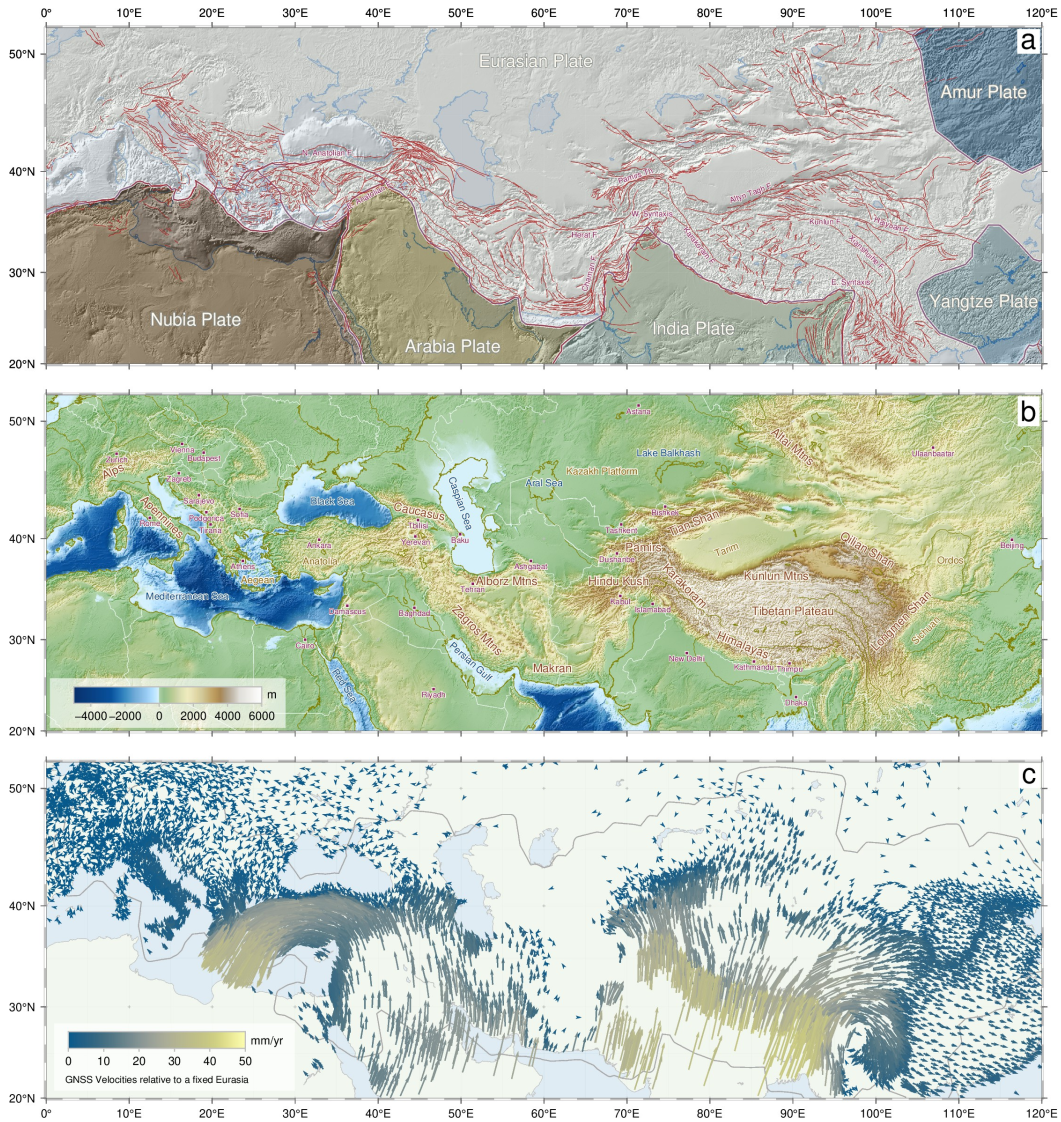

Figure 1: (a) Major active faults of the Alpine Himalayan orogeny (Styron and Pagani, 2020) with the tectonic plates of the region outlined from Bird (2003) and the major faults (F.) named in the text marked. (b) Topographic map of the belt with major mountain zones marked in brown and capital cities denoted by squares. (c) Compiled and aligned GNSS locations used in this study (Tables SA1–SA2) with velocities shown with respect to a fixed Eurasia reference frame (velocities with respect to the ITRF14 are shown in Figure SB1). The dark grey polygon outline denotes the area of coverage for the combined InSAR-GNSS velocity field generated here.

the seismic moment accumulation rate as this sets the budget for future earthquakes; geodetic strain-rate observations allow us to estimate this accumulation rate (Guns et al., 2024). At a crustal scale, a long-standing argument has been determining the most parsimonious way of describing the kinematic behaviour of the continental crust in terms of block-like behaviour (Thatcher, 2007), continuum behaviour (Houseman and England, 1986), or a mixture of both (Wang and Barbot, 2023; Fang et al., 2024a,b).

Using surface deformation over wide areas, we aim to understand processes occurring at depth in the lithosphere, where we cannot make direct observations. We can make sensible inferences of the behaviour of crustal material at depth in terms of the rock rheology (Bürgmann and Dresen, 2008) by examining time-dependent processes and testing possible spatial variabilities in strength and weaknesses of the lithosphere. We want to determine the forces involved in the collision of continents and during mountain building (Copley et al., 2010; Warners-Ruckstuhl et al., 2013; England et al., 2016; Fang et al., 2024a) and their evolution, erosion (Avouac and Burov, 1996) and subsequent gravitational collapse (Rey et al., 2001). Determining what level of influence the convecting mantle has on controlling the scale and pattern of surface deformation we observe today has also been a focus of research (Houseman and Molnar, 1997; Faccenna et al., 2021). We want to explain the patterns of fault frictional behaviour, and what promotes shallow aseismic creeping of active faults versus locking and strain accumulating (Harris, 2017; Jolivet and Frank, 2020). Finally, knowing whether the decadal geodetic rates of strain accumulation we measure today are consistent with the longer-term geological rates is important for understanding the evolution of fault zones, building models of the earthquake deformation cycle, and for enabling the use of short-term geodesy in assessing longer-term hazard (Thatcher, 2009; Meade et al., 2013; Elliott et al., 2016; Wright, 2016; Hussain et al., 2018; Mousavi et al., 2025).

There are also broader implications of our new data in terms of probing the link of tectonics to climate (Molnar and England, 1990), in particular understanding the growth of topography and evolution of geomorphology of mountains (Wolf et al., 2022) and the relative importance of surface processes and rheology. Also important is measuring and accounting for Glacial Isostatic Adjustment (Whitehouse, 2018), as well as constraining the distribution and behaviour of permafrost (Zwieback et al., 2024). Displacement time series of surface observations can provide information on water resources from the measurements of the widespread land subsidence and coastal subsidence (Galloway and Burbey, 2011), the latter of which has implications for relative sea-level rise (Wu et al., 2022). Additionally, such measurements, when at sufficiently high resolution, can be useful for volcano deformation monitoring (Ebmeier et al., 2018; Poland and Zebker, 2022), determining landslide stability (Cohen-Waeber et al., 2018; Bekaert et al., 2020) and for monitoring mining activities (Yang et al., 2020) and critical infrastructure (Wu et al., 2020).

To address these questions, we require a knowledge of the current rates of surface deformation and the degree of localisation of strain, as these provide key inputs for constraining future models and act as secular rates to compare against when exploring time-varying behaviour of crustal deformation. Previous studies have been able to make significant progress using lower density GNSS data (Reilinger et al., 2006; Thatcher, 2009; Stamps and Kreemer, 2024), or local and regional InSAR based studies (Elliott et al., 2016). The widespread availability of Sentinel-1 data (Torres et al., 2012) is a timely opportunity to produce an internally referenced and consistent continent-wide InSAR-derived velocity field. This is particularly important for areas that, unlike Europe (Piña-Valdés et al., 2022) and the Western U.S. (Herring et al., 2016), lack (and are unlikely to readily achieve) dense GNSS coverage.

In this paper we introduce the tectonics of the AHB and previous studies of regional scale deformation, before presenting the two datasets we are principally using – GNSS and InSAR. We describe our compilation of previously-published GNSS velocities and how we align them to create a consistently referenced set of data. We then show our methodological and systematic approach to generating and correcting the InSAR time series before illustrating how we integrate this with the GNSS constraints to generate velocity and strain rate fields at a trans-continental scale. We show the resulting components of velocity and measures of strain across the Alpine Himalayan Belt as a whole, before focusing on a few key deforming tectonic areas. We discuss the sources of these signals and the limitations of measuring strain in the presence of the remaining noise in our time series. We finish by discussing the potential implications for our broader understanding of continental deformation and the outlook for improvements expected with data from future satellite missions.

2. Background

The AHB has long been recognised as a wide collisional zone (Argand, 1922) and exhibits a range of deformation styles running from west-to-east (Jackson and McKenzie, 1984). There are three major sections of the collisional zone that predominantly align with the three converging plates to the south of Eurasia (Figure 1). The tectonics of the Mediterranean are dominated by the relative motion of Africa (Nubia Plate) and the rate of convergence has slowed over the past few tens of millions of years from > 30 mm/yr before 25 Ma to 13 mm/yr for 11–25 Ma and most recently < 7 mm/yr in the last 11 Ma (McQuarrie et al., 2003; Reilinger and McClusky, 2011). The deformation of the Turkish Iranian Plateau, Iran and the Makran is controlled by the Arabian plate with its northward motion (and anticlockwise rotation). The rates of relative convergence have slowed less significantly than for the case of Nubia-Eurasia, from the same > 30 mm/yr before 25 Ma to 21 mm/yr for the past 25 million years (Reilinger and McClusky, 2011). Finally, in the East, the Himalayas, Pamirs, Tibetan Plateau and Tian Shan are all controlled by the relatively rapid northward motion of India which has slowed from 58–83 mm/yr between 45–20 Ma to 44–57 mm/yr for the period 11–20 Ma and now to a rate of 34–44 mm/yr for the past 11 Ma (Molnar and Stock, 2009). In each of these, the distribution of gravitational potential energy of the regions modifies the pattern of deformation imposed by the boundary forces (England and Houseman, 1989; England and Molnar, 2015; England et al., 2016; Walters et al., 2017; Fang et al., 2024a).

2.1. Remotely Sensed Data and Technological Advances

Our ability to determine the deformation of the continents has been greatly advanced in the past two decades by space-based technologies and data sets (Elliott et al., 2016;

Stamps and Kreemer, 2024). Initial progress was made through regional GNSS studies (Feigl et al., 1993; Abdrakhmatov et al., 1996; Reilinger et al., 1997; Wang et al., 2001; Calais et al., 2003; Nocquet and Calais, 2003), before the global strain rate model (GSRM-1) provided horizontal velocities and low-resolution strain rates for most of the deforming plate boundary zones from over 3,000 GNSS sites globally (Kreemer et al., 2003). This was advanced in version 2.1 of GSRM (Kreemer et al., 2014) by incorporating over 22,000 velocities in the model. The most recent Europe Velocity model for 2022 (EuVeM2022) is constrained by data from 2,549 GNSS sites with an output resolution of 0.1° (Steffen et al., 2025) and covers Europe and Anatolia.

Distributed measurements of surface velocities from InSAR over deforming regions, when suitably referenced with GNSS data, offer the potential to advance our understanding of continental deformation by increasing the resolution of velocity fields (and the strain rates that can be calculated from them) where GNSS coverage is sparse. Over the course of the past quarter of a century, there has been a transition from the earliest studies with InSAR on slowly deforming regions solely focusing on single tracks of data less than 100 km in extent (Peltzer et al., 2001; Wright et al., 2001), largely with ERS-1/2 data (Bürgmann et al., 2000), to entire fault systems (Tong et al., 2013; Hussain et al., 2018) predominately with ENVISAT data, and more recently using Sentinel-1 SAR to capture deformation over large tectonically deforming regions (Xu et al., 2021; Lemrabet et al., 2023; Fang et al., 2024a) and whole country scales (Dehls et al., 2019; Drouin and Sigmundsson, 2019; Weiss et al., 2020; Festa et al., 2022; Zhang et al., 2024), and now large fractions of continents (Costantini et al., 2021).

As well as the technical advancements in the SAR instruments themselves, typified by the changes from ERS-1/2 (Attema et al., 1998) through ENVISAT-ASAR (Desnos et al., 2000) to Sentinel-1A/B (Torres et al., 2017), enabling more systematic acquisitions and observations suitable for interferometry, there have been improvements in processing SAR (Goldstein and Werner, 1998; Rosen et al., 2012; Sandwell et al., 2011), mitigating atmospheric noise (Lohman and Simons, 2005; Li et al., 2005; Jolivet et al., 2014; Bekaert et al., 2015) and enhancing time series analysis (Hooper et al., 2004;

Hetland et al., 2012). This has enabled ever smaller displacement rate signals (mm/yr) to be imaged over large spatial scales of many hundreds of kilometres. These improvements include accounting for the tropospheric variability in delay using weather models (Jolivet et al., 2011; Murray et al., 2019) and the ionosphere (Liang et al., 2019) using global ionospheric maps, enabling smaller signals over longer wavelengths to be discernable.

The large data volumes have also necessitated a change in approach as the input data alone is PetaByte scale, with automation of processing of Big Data needed in processing pipelines on High Performance Computing (HPC), as used here (Lazecký et al., 2020). Various groups around the world have initiatives for processing areas or providing on-demand processing to provide analysis-ready data, enabling wider adoption by the broader community without the need for large computing power. ForM@Ter LArge-scale multi-Temporal Sentinel-1 InterferoMetry (FLATSIM, Doin et al., 2011) is an initiative providing processed products over broad regions for research. The Alaska SAR Facility (ASF) archives the SLC data from ESA and provides processing capability (Meyer et al., 2025). They also host the Jet Propulsion Laboratory (JPL) geocoded and unwrapped interferograms (Bekaert et al., 2023). The European Ground Motion Service is available for (most of) Europe (Costantini et al., 2021) with high-resolution velocities available, but with coverage limited to Europe. The German Aerospace Center (DLR) has a pipeline for processing all of the deforming regions of the globe (Gomba et al., 2024). Efforts are also being made to make InSAR data easier to use in the form of Analysis Ready Data (Wang et al., 2022b).

2.2. Challenges and Approach Taken Here

There are a number of challenges to generating continental-scale velocity and strain rate fields given the large spatial scale (millions of square kilometres) and fine sensitivity sought (mm/yr velocities and strain rates of tens of nanostrain per year). The volumes of data and their ongoing acquisition, whilst very welcome for science, present constant data handling and processing challenges. Whilst Sentinel-1 SLC (Single Look Complex) data has pixel spacings of $\sim$4 m in ground range and 14 m in azimuth, it has been necessary to multilook the interferometric images to 50 m and downsample further to 100 m as a compromise to facilitate processing and storage. When working with the time series analysis and velocity field generation, further downsampling has been necessary, giving a spatial scale of 1 km in the data presented here to make the processing tractable.

Noise from the troposphere can be significantly reduced by applying weather model corrections (Bekaert et al., 2015; Murray et al., 2019), except in areas of steep topography. The assessment of corrections for the ionosphere over long wavelengths is still developing, as the variability of the total electron content, which advances the phase of the signal, depends on latitude and time of acquisition (Liang et al., 2019). Being C-band, Sentinel-1 is less susceptible to ionospheric phase variability than L-band satellites, but the signal is still significant. The 11-year solar cycle (Hathaway, 2015) has a large impact on the ionosphere (Roma-Dollase et al., 2018); the solar cycle 24 high occurred around the launch of Sentinel-1A in early 2014, and solar activity was therefore low for the early part of the time series in this study (starting early 2016), but it has been increasing and was more elevated than expected by our end date in early 2024 (Figure SB7).

Whilst shorter wavelength C-band is better in terms of ionospheric noise, it is less good in terms of phase coherence than L-band, leading to a loss of data coverage in some areas, in particular those that are more vegetated and cultivated, where surface properties change rapidly (Kellndorfer et al., 2022). This affects the AHB most markedly in the west (Europe to Anatolia) and in the south-east (around the Himalayas and Eastern Tibet). This is in part mitigated by the short revisit time of Sentinel-1, which was particularly good when S1A and S1B were both in operation, up until the loss of the SAR antenna power supply unit on S1B on 23rd December 2021. Issues of coherence were further compounded when there were large acquisition gaps (sometimes year-long gaps) over certain regions due to limitations of the duty cycle. Additionally, data was not acquired in Eastern China (Figure SB2). High mountain areas with steep relief such as the Caucasus and Himalayas (Grandin et al., 2012) lead to further gaps in data coverage and large phase unwrapping errors due to the loss of connected coherence across our scenes.

Interferometric coherence with Sentinel-1 has been better maintained (Figure SB9), when compared to ENVISAT and ERS-1/2, by improved orbital control. The orbital tube target originally targeted ~ 100 m (Prats-Iraola et al., 2015), keeping it well within the critical baseline for interferometry. However, recent reduced orbital inclination control on S1A, from partial failure of the thruster for out-of-plane manoeuvres in February 2024, has led to increasingly large perpendicular baselines (Pinjeiro, 2024) that are likely to have a detrimental impact on interferometric performance and capabilities.

InSAR data, when used alone, is severely hampered by providing a single line-of-sight and only relative deformation (Hu et al., 2014). Even when two look directions are combined, polar-orbiting satellites that look to one side only are largely insensitive to north-south motion (Wright et al., 2004b). North-south information can be extracted from burst overlap areas Sentinel-1 (Li et al., 2021a), but the coverage is poor. Due to this lack of 3-component information and the need to reference the InSAR data, we also use GNSS data. There are significant numbers of GNSS sites in Europe and Eastern China, but there is a low density of sites in the centre of the AHB around Afghanistan and Pakistan (Figure SB1). This presents challenges when trying to reference the InSAR data and interpret long-wavelength deformation patterns.

Our aim is to measure surface displacements and calculate velocities that are representative of tectonic processes due to deep-seated deformation sources of stress causing strain over long wavelengths. However, near-surface processes such as land subsidence (Haghshenas Haghighi and Motagh, 2024; Wu et al., 2022) and permafrost (Chen et al., 2022; Fan et al., 2025) mask deeper-seated tectonic strains, in particular in regions where we would want to measure uplift rates reliably, and therefore these areas must be interpreted carefully.

We derive decadal surface velocity fields from which we calculate crustal horizontal strain rates. Our aim is to capture the long-term, interseismic part of the earthquake cycle (Wright, 2002) in these measurements. However, the velocity field is calculated over a finite duration and incorporates the coseismic offsets of recent earthquakes, which we need to remove from our time series as they otherwise manifest as large strains due to the elastic release of strain accumulation. Furthermore, the stress from the earthquakes themselves results in postseismic deformation that can be attributed to a number of causes (Wright et al., 2013; Ingleby and Wright, 2017) and lasts for many years and potentially decades. Given that the earth behaves viscoelastically, and faults can slip aseismically, it is not straightforward to attribute interseismic strain accumulation to fault slip potential (Hussain et al., 2018), depending on the rheology of the fault zone and crust (Hilley et al., 2009) and when within the earthquake cycle the deformation is measured (Savage and Prescott, 1978; Wang et al., 2021). A knowledge of historical earthquake locations enables us to test to what degree the elevated strains we are measuring today result from the postseismic deformation of past seismicity.

Sentinel-1 presents relatively moderate resolution SAR data at the 5–20 m scale. However, it is necessary to work with multi-looked data for increased spatial correlation, producing pixels of equal dimensions and reducing data volumes in the Small Baseline Subset (SBAS) process. However, the process of multi-looking of interferograms results in mis-closure of the phase loops because the filtering effect is non-conservative for some land cover types (De Zan et al., 2015). This is attributed to the vegetation and soil water content (De Zan et al., 2014; De Zan and Gomba, 2018). In regions of agriculture, careful interpretation of the apparent ground deformation is required (Jiang and Lohman, 2021) as these regions result in large signals varying in space and time if cultivated. Soil moisture effects have also been shown to have an effect on the phase closure in arid regions (Lohman and Bürgi, 2023). This issue is exacerbated for short-period, multi-looked interferograms such as used here that result in a fading bias (Ansari et al., 2021), which is now prevalent with the shorter revisit time of 1–2 weeks offered by Sentinel-1 when compared to ENVISAT and ERS-1/2 acquisitions that were typically at least spanning a month. There are empirical mitigation approaches that may be possible to apply at scale in future (Maghsoudi et al., 2022).

Some of these challenges have been addressed here with our processing strategies and this study provides an important historical record of recent deformation and a baseline for change detection in upcoming decades. However, future upcoming missions should also provide improvements in these areas. The recently launched Sentinel-1C satellite, and planned missions such as NISAR (Kellogg et al., 2020), Harmony (Kääb et al., 2024), ROSE-L (Davidson and Furnell, 2021), and Sentinel NG (Torres et al., 2024) will be discussed later in this context.

2.2.1. Aims and Objectives

Our aim is to produce a dataset that is useful to inform ideas of continental deformation and aid in the improved assessment of seismic hazard. Our objectives are to produce a compilation of GNSS velocities for the AHB and use these with InSAR data to create a geodetically constrained displacement time series and average velocity field for the region that is useful to the wider community. Our approach has been to build on the individual areas in the Alpine Himalayan Belt for which we have already produced velocity fields using Sentinel-1, covering from Türkiye (Weiss et al., 2020) through Iran (Dodds et al., 2022; Watson et al., 2024), the Tian Shan and onto the Tibetan Plateau (Ou et al., 2022; Wright et al., 2023; Shen et al., 2024a; Wang et al., 2024; Fang et al., 2024a). We have updated the data in time, and greatly expanded the spatial coverage to unify them into a single velocity field, by performing near-uniform interferometric processing with common time spans where possible to bring all the regions up-to-date in a consistent manner. This is then tied to the new compilation of GNSS velocities we generate here for the Eurasian plate to provide a fully referenced dataset.

3. Data

Our velocity field is derived from the combination of two key geodetic observations of the recent displacement history of the Earth's surface: 1) GNSS and other point velocities, which provide a comparatively sparse picture of surface motion within an internally consistent reference system; and 2) InSAR time series of distributed surface displacements, which provide dense estimates of relative velocity across broad areas, but which need to be suitably referenced to reliably interpret over longer wavelengths.

3.1. GNSS and other point velocities

To underpin the InSAR velocities and tie them to a unified reference frame, we compile 58,970 point velocities across the Alpine-Himalayan Belt (Figure SB3) from 150 studies and datasets. This includes 49,608 GNSS velocities and 9,258 levelling rates, plus 20 DORIS, 66 SLR and 18 VLBI velocities. Some of the 150 source datasets (Tables SA1–SA2) are themselves compilations of datasets, such as the compilation that partially underpinned the Global Strain Rate Model 2.1 (GSRM) (Kreemer et al., 2014), so in total there are 268 original datasets. We prioritise velocities that are at least partially based on measurements from the Sentinel-1 era (2014–present), so our approach is to compile the GSRM compilation (pre-2014 data) plus every findable and suitable velocity dataset published since then, and to replace pre-2014 with post-2014 velocities where suitable. We do include a few datasets that predate the GSRM; almost all of these provide vertical motion rates that we consider high-quality (the GSRM used only horizontal velocities). We exclude several (otherwise high-quality) velocity fields that include significant postseismic deformation from the 2008 Wenchuan, 2005 Kashmir, or other earthquakes that predate the Sentinel-1 era by several years. Section 4.1 describes the process of converting this compilation into a usable picture of continental surface deformation.

3.2. SAR

We use Single Look Complex (SLC) data from the Sentinel-1 C-band SAR instrument (centre frequency 5.405 GHz) with VV polarisation acquired by the European Space Agency (ESA). We use these data to derive surface displacement measurements across the Alpine Himalayan Belt using differential interferometry (Section 4.2). The Sentinel-1 mission provides systematic near-global radar imaging (Torres et al., 2012; Potin et al., 2016), enabling precise monitoring of long-term ground deformation at a continental scale. Most of the tectonic belts are imaged with both ascending and descending orbits, providing two independent viewing geometries that enable us to separate east-west and vertical deformation.

We source the data from both the Alaska SAR Facility (Meyer et al., 2025) and also through the Near Line Archive at UK's data analysis facility for environmental science (JASMIN). The data we select typically span from early 2016 (beginning of March) to early 2024 (end of February, see Supplementary Material Appendix C for exact dates) except when we choose to start the time series late or finish it early to mitigate the effects of earthquakes (Section 4.4). The constellation comprises both Sentinel-1A and its near identical pair Sentinel-1B (Miranda et al., 2017) during much of this time. The AHB from southern Europe to eastern China is captured by 56 partially-overlapping, 250-km-wide orbital tracks in each of the ascending and descending directions (Figure SB2). At lower latitudinal areas in our study (25°N), the tracks overlap by about two-thirds of a subswath (∼50 km), and at the highest latitude (50°N) by over one subswath (∼100 km). This yields complete spatial coverage and some redundancy in each direction, except at the edges of the polygon shown on Figure 1 C, and which we use for velocity uncertainty analysis (Section 4.7). Data are typically available in both look directions, enabling us to resolve east-west and vertical components of motion with the use of GNSS to constrain north-south motions (e.g., Wright et al., 2004b; Weiss et al., 2020). An exception is the easternmost part where there is a gap in descending data between the Eastern Tibetan Plateau and Beijing around the Ordos Plateau (Figures 1 & SB2).

We divide the entire AHB into 844 LiCSAR (Looking inside the Continents from Space) frames (Lazecký et al., 2020), consisting of 436 ascending and 408 descending frames (Figure SB2). Each frame comprises multiple bursts (typically 13 in each of the 3 subswaths) acquired in Interferometric Wide Swath (IW) mode (Geudtner et al., 2014) using Terrain Observation by Progressive Scans (TOPS) (Yagüe-Martínez et al., 2016). The total dataset includes 221,569 Sentinel-1 epochs, providing a dense spatio-temporal sampling of surface displacement across the region. From 2016 to 2021, the region from southern Europe to eastern Türkiye was covered by 6-day repeat-pass acquisitions, benefiting from the combined operation of Sentinel-1A and Sentinel-1B. However, following the failure of Sentinel-1B in December 2021, the acquisition frequency for this region was reduced to 12 days. For the rest of the study area, including the central and eastern portions of the AHB, the 12-day repeat cycle was used throughout the observation period where possible, although some tracks experience longer revisit times and temporal data gaps.

3.3. Ancillary data

In addition to the SAR datasets themselves, a number of ancillary datasets are needed in the processing chain, in particular to apply corrections to the interferograms for phase changes resulting from the changing atmosphere, including phase delays through the troposphere (Yu et al., 2018b), and phase advances through the ionosphere (Yunjun et al., 2022). Details of these corrections are provided in Section 4.3.

We also require precise orbital ephemerides (POE) data for satellite positions to facilitate the coregistration steps between SLCs, and use the precise orbits provided by ESA (Fernández et al., 2024). In the InSAR processing, it is also necessary to account for topography in the simulated phase. Additionally, a geographic dataset is required to reference the interferograms from radar to the geographic coordinate system. We use a mix of both the SRTM (Farr et al., 2007) and Copernicus GLO-30 (Rizzoli et al., 2017) Digital Elevation Models (DEMs) to achieve this, which is recorded in the metadata for each frame.

4. Methodology

Our approach starts with compiling a consistent set of GNSS velocities from previously published studies over our area of interest. We then create line-of-sight velocities and time series from ESA Sentinel-1 satellite radar data through interferometric processing. The InSAR velocities are combined with the aligned GNSS velocities to create surface velocities in a Eurasian reference frame and strain rates as output datasets for further interpretation (Figure 2).

4.1. GNSS: generating a reference frame

The preparation of the GNSS and other pointwise velocities is as follows. We first transform each individual velocity field from its published reference frame into

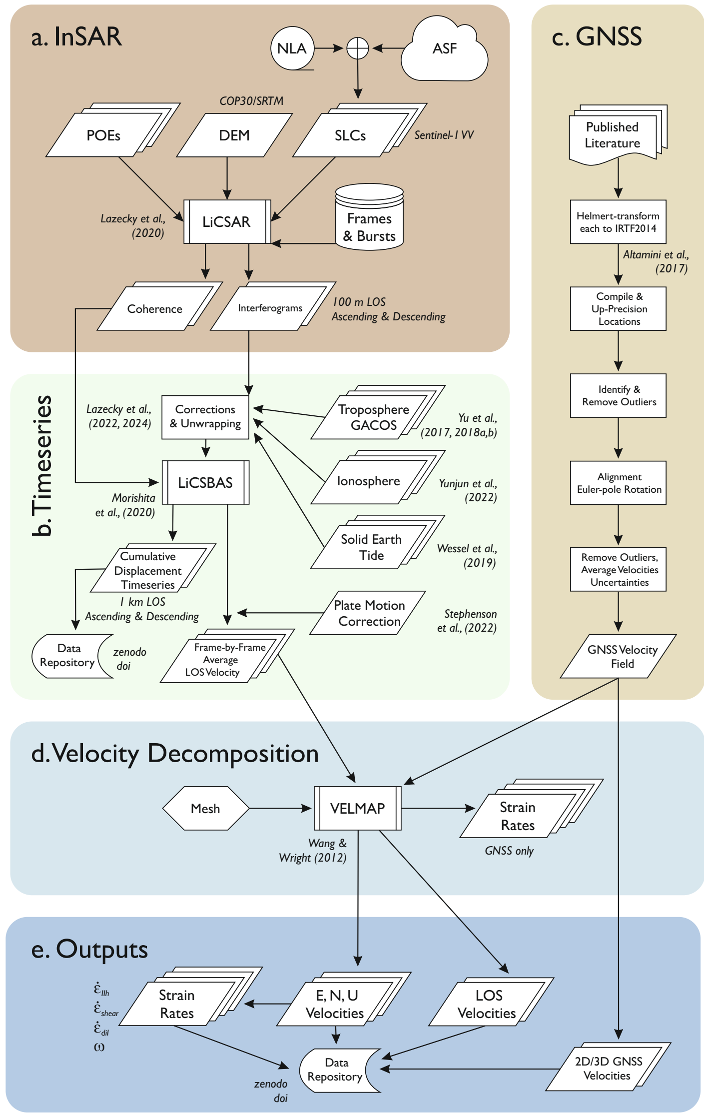

Figure 2: Workflow for (a) InSAR processing chain and (b) correction and time series calculation, (c) GNSS compilation and alignment, (d) unified velocity generation and (e) final strain rate field calculations and outputs. InSAR - Interferometric Synthetic Aperture Radar, NLA - Near-Line Archive, ASF - Alaska SAR Facility, POE - Precise Orbital Ephemerides, DEM - Digital Elevation Model, COP30 - Copernicus 30 m, SRTM - Shuttle Radar Topography Mission, SLC - Single Look Complex, VV - Vertical Vertical Single Polarisation, LiCSAR - Looking inside the Continents from Space SAR, LOS - Line-of-Sight, LiCSBAS - Looking inside the Continents from Space Small Baseline Subset, GACOS - Generic Atmospheric Correction Online Service, GNSS - Global Navigation Satellite System, ITRF2014 - International Terrestrial Reference Frame, E, N, U - East North, Up. $\dot{\varepsilon}_{II_h}$ = second invariant of strain rate, $\dot{\varepsilon}_{shear}$ = maximum shear strain rate, $\dot{\varepsilon}_{dil}$ = horizontal dilatation rate, ω = vorticity, (Savage et al., 2001).

ITRF2014 (Altamimi et al., 2017) using the *itrstrafo* tool in Geodetic Transformations (`https://www.mathworks.com/matlabcentral/fileexchange/9696-geodetic-transformations`). (The one exception is the levelling dataset from Jackson and Bilham (1994) sourced through Dal Zilio et al. (2020)). This puts all of the velocity fields nominally in ITRF2014; however, different studies used different global or regional velocities to get into their original reference frames, and so some systematic differences remain between these velocity fields, which we correct in a later step.

Next, we "up-precision" the station locations of the GNSS velocities (not the leveling, DORIS, SLR or VLBI) by first collecting different studies' locations for each station (presented in geographic coordinates of decimal latitude and longitude). This is achieved by finding matching station names and published locations within 2 km of each other (or within 20 km if one study rounds locations to the nearest 0.1°. These search distances worked best out of a number of options tested). For each station, if any studies provide the location to four or more digits after the decimal point (i.e. to a precision of 10 m), we use the median of those locations as the station's location across all studies. If none of the locations are that precise, we use the median of locations provided to three digits after the decimal point; otherwise, we take the median of all available locations excluding studies that categorically round to 0.1°. We then rotate all of the velocities into the ITRF2014 Eurasia-fixed reference frame (Altamimi et al., 2017). The dataset at this stage is shown in Figure SB3.

Next, we identify 74 velocities for which the original study does not provide an uncertainty or gives an uncertainty of zero (on one or more components of velocity) but does indicate the total occupation time and number of occupations at the site. We use these metadata to estimate the uncertainties on these velocities assuming a model of combined random-walk, white and flicker noise. For random-walk noise, we use horizontal and vertical coefficients from Castro-Perdomo et al. (2022) and the equation from Beavan et al. (2016). For white noise, we we average several of the lower-end coefficients provided by Williams et al. (2004) and Wang et al. (2012) and use the equation from (Beavan et al., 2016); for flicker noise, we average several lower-end coefficients from

Williams et al. (2004) and Wang et al. (2012) and use the equations from Williams (2003) and (Bos et al., 2008), averaging the results from the two equations.

Next, we find any very small positive velocity uncertainties and round them up to a minimum of 0.05 mm/yr uncertainty for east and north components (the 2.275th-percentile or -2σ value of all assembled east and north uncertainties) and 0.12 mm/yr uncertainty for the vertical component (the -2σ value of all assembled vertical uncertainties).

Next, we begin earmarking and removing outlier velocities. We first earmark 739 velocities that have site occupation times < 1.5 yr, have east or north uncertainties of > 5 mm/yr or vertical uncertainties of > 24 mm/yr (the $+3\sigma$ values of the distributions of those uncertainties), or have components with unresolved zero or non-existent uncertainties. Many GNSS studies use 2.5 yr as a lower-bound occupation time for including velocities (e.g., (Bock et al., 2021), (Gorshkov et al., 2023)), but there are isolated velocities with $<$ 2.5-yr occupation times in remote areas such as Afghanistan that we were motivated to retain. We also earmark 228 additional velocities that we deem problematic but which are not automatically flagged by the subsequent outlier-detection method. This includes, for example, the velocities at sites CHBR and SRVN from Frohling and Szeliga (2016) and Khorrami et al. (2019), which Abbasi et al. (2023) showed were likely affected by offsets. It also includes velocities from Mahmoud et al. (2013) in southeast Türkiye included in Kreemer et al. (2014) and Palano et al. (2018); these are difficult to reconcile with nearby velocities from other studies and have comparatively large uncertainties.

We then use a bespoke algorithm to automatically detect and remove other outlier velocities, as follows:

- For each velocity V_i with horizontal (east and north) components, we go to its location and "scan the horizon" in eight directions (N, NE, E, SE, S, SW, W, NW). We select the nearest X (here $X = 7$) velocities in each direction (within a 180-degree-wide "aperture" centered on each direction), excluding velocities already earmarked as outliers, those that do not include horizontal components, those more than within

2° (222 km) distant, and V_i itself. We combine all of these nearby velocities into a population (separating out their east and north components), henceforth called a "panorama".

- The choice of X= 7 was arrived at through testing, but it does not substantially affect the results unless changed significantly; the overall goal is to have an adaptive spatial kernel that usually comprises a few dozen velocities to work with for robust statistics.

- If the panorama contains at least X (7) total velocities, we compute their median in each horizontal component and their mean absolute deviation (MAD) about that median. We also compute their mean and the MAD around that in each horizontal component, as well as the average of the mean and median and the MAD around that average, to reduce the impact of the specific choice of central value.

- If – in either the east or north component and using any of those metrics – velocity V_i differs from the panorama's central value by more than the corresponding MAD times a scalar multiple (called "MADmult"), we mark Vi as an outlier.

- This runs for several iterations, over which MADmult is started at a relatively large value to flag the largest outliers first, and is gradually brought down to a minimum value with more outliers progressively being earmarked (and thereafter excluded from any panorama). Through testing, we set the minimum value of MADmult to 3.75. In brief, we define outliers as east or north velocities that are at least 3.75 mean absolute deviations outside the central value of nearby velocities.

By the end of this step, 3,386 velocities are earmarked as outliers and are then removed (Figure SB4).

Next, we align the velocity datasets (minimise systematic differences between them) using an iterative approach. For each study S_i, we tabulate the differences in horizontal velocities between it and all other studies at common sites. We also compute the formal uncertainties on these differences (the 2-norms of the individual velocity uncertainties in each pair of velocities being differenced). Then, in each of 5,000 realisations (for S_i), we extract a random-length subset of these horizontal velocity differences, perturb those differences by their corresponding uncertainties times a Gaussian-random vector, and least-squares fit the perturbed velocity differences to a rotation about an Euler pole. If the azimuths of velocities in the least-squares solution vary by more than 90° across the field, we discard the solution; otherwise, we save it and proceed to the next realisation. At the end of the 5,000 realisations, we take the median of the (non-discarded) solutions as the preferred correction to be applied to study S_i's velocity field. We compute the uncertainty on this median correction as 1.2533 times the standard deviation of the non-discarded solutions (https://influentialpoints.com/Training/standard_error_of_median.htm).

However, we apply only 1/20 of the correction to S_i. We then save the modified velocity field as S_{i-mod} and proceed to the next study S_j. When correcting S_j, we use the pre-modified version of S_i (and all other studies), so that all studies are corrected "simultaneously" and their order does not matter. After running this on all studies, we update all of the studies' velocities and uncertainties simultaneously, and recompute their horizontal velocity differences and the uncertainties on those differences. We run this procedure over 100 iterations. The choice of applying 1/20 of the correction at a time guarantees stable convergence over 100 iterations; we note that correcting 1/20 of a difference 100 times is in principle equivalent to correcting 0.994 of the original difference.

We also run two overall sets of 100 iterations. In the first 100, we treat compilation studies as single datasets (equivalent to assuming that the compilations already aligned their constituent studies adequately). In the second 100 iterations, we split out the compilation studies into their constituent datasets and treat all 268 datasets as independent. This minimises any remaining systematic differences that may have arisen from different compilations' use of varying alignment methods. The one exception is that we split out the GSRM compilation (Kreemer et al., 2014) into its constituent datasets from the beginning; this is because the velocity field of Aktuğ et al. (2013) in central Anatolia is visibly deflected southward compared to other datasets within the GSRM and we aim to correct that as early as possible. The aligned velocity field is shown in Figure SB5.

After aligning the velocity fields, we rerun the automatic outlier detector (with the same settings as previously) and remove 362 additional velocities that now stand out amongst the aligned velocity field enough to be earmarked as outliers. Next, we remove older velocities that are superseded by high-quality newer velocities. Out of the 268 datasets, we identify 145 that are entirely or mostly based on pre-Sentinel-1-era measurements; the majority of these are from Kreemer et al. (2014). We compile a separate list of 27 newer studies (providing horizontal and/or 3D velocities) that we consider high-quality. From the 145 older datasets, we identify 5,705 velocities that are collocated with (located $\leq$ 2 km from) a velocity in one of the 27 preferred studies, and remove the older velocities. In addition, 12 of the 27 preferred studies provide 3D velocities, and we use these to supersede 3D velocities from less preferred sources, removing 2,515 additional velocities this way. Finally, we identify and remove 1,468 remaining duplicates (cases where one dataset has been used by multiple studies). This leaves 45,532 of the original 58,970 velocities, including 32,608 horizontal and 24,592 vertical velocities (Figure SB6).

Finally, we combine collocated GNSS velocities (not the levelling, DORIS, or SLR, which we leave separate). At locations with multiple GNSS velocities, we take the combined velocity in each component (east, north and up) to be halfway between the mean and the median of the available velocities in that component. In estimating the uncertainties on the combined velocities, we iterate over whether or not to incorporate the uncertainties on the individual velocities, whether to treat different velocities being combined at a site as independent data, and which metric of spread to use. For the unweighted mean of N velocities being combined (the first half of the central value), we consider eight alternative measures of the uncertainty on the mean:

1. the standard error (the standard deviation of the input velocities divided by $\sqrt{N}$);

2. the standard deviation of the input velocities (equivalent to setting N = 1 in the $\sqrt{N}$ term, or roughly assuming that the studies are not independent);

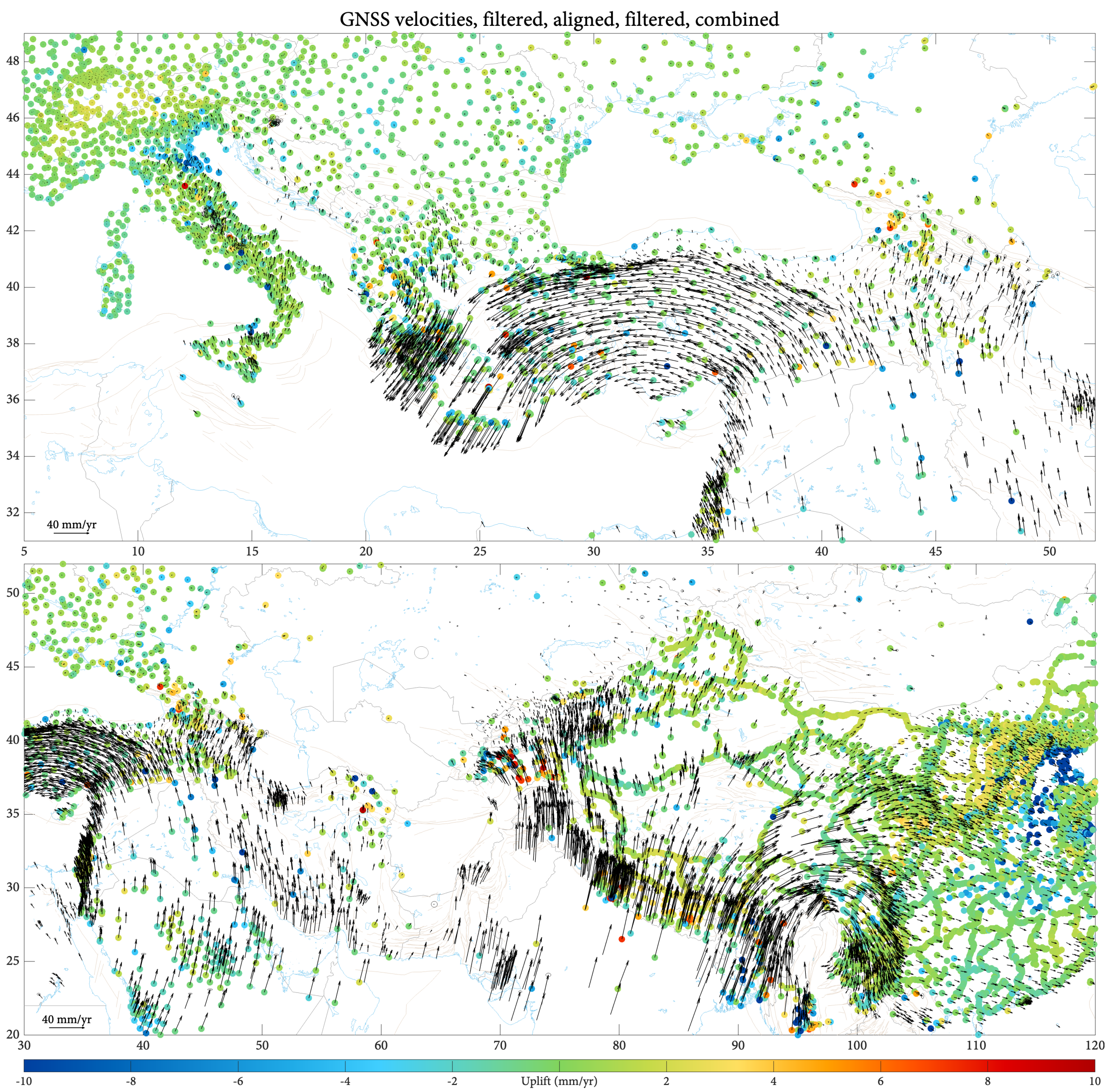

Figure 3: The prepared dataset of GNSS and point velocities after removal of outliers, Euler-pole alignment of studies, removal of superseded older velocities and duplicates, and combination of collocated velocities, in the western AHB (top) and eastern AHB (bottom). Arrows show horizontal motions with respect to the ITRF2014 Eurasia-fixed reference frame (Altamimi et al., 2017); coloured circles show vertical motion rates.

3. the standard error of the multimodal PDF that can be constructed from the velocities and their individual uncertainties (assuming the uncertainties are Gaussian);

4. the standard deviation of the same multimodal PDF;

5. the mean absolute deviation (about the mean) of the input velocities, multiplied by 1.2533 (Geary, 1935);

6. the MAD (about the mean) of the input velocities, multiplied by 1.2533, divided by $\sqrt{N}$;

7. the MAD (about the mean) of the multimodal PDF, multiplied by 1.2533;

8. the MAD (about the mean) of the multimodal PDF, multiplied by 1.2533, divided by $\sqrt{N}$.

For the unweighted median (the second half of the central value), there are eight similar alternative measures; the first four terms above are multiplied by 1.2533 and the last four use the median as the central value rather than the mean. We compute these eight measures of uncertainty for each part of the central value and compute the 2-norm of every possible pair of uncertainties between the sets (each divided by 2), for a total of 64 alternative estimates of the uncertainty. We take halfway between the mean and median of these 64 values as the preferred uncertainty on each component of each combined velocity.

At the end of this combination, there are 19,608 point velocities spanning the Alpine-Himalayan Belt, including 6,064 3D velocities, 4,424 2D velocities and 9,109 vertical rates (Figure 3). In total, 10,488 velocities provide constraints on horizontal surface motion (i.e. are 2D or 3D) and 15,173 constrain vertical motion (are 3D or pure vertical). The spatial footprint of each input study, the number of velocities it provides, and the number of velocities that are ultimately used are given in Tables SA1–SA2. Note that in the subsequent modelling, we exclude the purely vertical velocities (most of which are inherited from levelling measurements) and vertical rates with magnitudes of > 20 mm/yr.

4.2. LiCSAR: from SAR to InSAR

We follow the methodology already outlined in Lazecký et al. (2020), starting from Level-1 Single Look Complex data from Sentinel-1A/B to generate a redundant network of over 1,100,000 interferograms over the Alpine-Himalayan Belt. The interferograms are multilooked and geocoded to grids of 0.001 degrees (approx. 110 m) resolution in the WGS-84 reference system. As well as the standard four consecutive interferometric connections (forwards and backwards) per temporal epoch (same for both 6 or 12 days revisit time), we densify the network forming additional combinations that cover longer time periods of up to 12 months for some regions such as the Tian Shan. Such network including long-span interferograms helps overcome limitations such as data gaps due to decorrelation resulting from the presence of snow and a fading phase bias that can affect short-span interferograms (Maghsoudi et al., 2022).

The methodology generating the InSAR dataset was further developed and some of the geographic units (frames) use different ancillary data such as the DEM and orbit ephemerides, or a different methodology in phase unwrapping and co-registration where we dropped the original intensity cross-correlation operation in azimuth direction to prevent errors during significant ionosphere (Chen and Zebker, 2014) as one of the factors causing a two-fold increase in azimuth errors that we observed in data prior to 2016 (Lazecký et al., 2023). We provide basic metadata in the repository datasets on applied DEM per frame, but other differences are not flagged as they are considered negligible for standard InSAR measurements.

In some circumstances of reusing updated coregistration lookup tables with different versions of ephemerides, we introduced azimuth offset errors in some data, manifesting as burst discontinuities including in the final velocities if left uncorrected. This affects a small fraction of interferograms and we identify and skip such interferograms based on outlying azimuth offsets per epoch. We ensure the consistency of the data by selecting only data since March 2016 up to March 2024, and performing an updated phase unwrapping routine (Lazecky et al., 2022).

The phase unwrapping utilizes up-to-date version of SNAPHU at its core (Chen and

Zebker, 2002) and is performed on the original LiCSAR geocoded unfiltered interferograms that are further multilooked by a factor of 10 in both latitude and longitude, leading towards the final resolution of 0.01×0.01 degree (approximately 1×1 km), and corrected for some of the non-tectonic signals from the troposphere, solid Earth tides, and ionosphere (see Section 4.3), where the line-of-sight correction maps were generated per epoch and combined to form simulated correction interferograms. Removal of these terms improves the quality of the unwrapping procedure as observed in LiCSBAS (Looking Inside the Continents from Space Small Baseline Subset) quality metrics (Morishita et al., 2020) described later. For a portion of frames, the unwrapped data with coherence estimated below 0.15 were masked prior to the time series inversion, effectively avoiding the propagation of errors due to noise into the inversion. The processing parameters were optimised for each individual frame, with parameters listed in Supplementary Materials Appendix C.

4.3. Corrections of atmospheric and tidal signals in interferograms

A number of corrections are made on an epoch-by-epoch basis to reduce the noise (Figure 2). This includes phase delays/advances from both the troposphere and ionosphere and non-tectonic surface deformation from solid earth tides, which we can model and remove. An analysis of average correction gradients shows that the effect of the troposphere is an order of magnitude higher than the other terms (Figures SB7 and SB8). The velocities inverted from ionosphere corrections also display significant quasi-planar ramps for ascending frames, which are acquired at around 6 p.m. local time. This is demonstrated for an example at a lower latitude in Figure 4.

4.3.1. Troposphere

We use the tropospheric corrections derived by the Generic Atmospheric Correction Online Service for InSAR (GACOS, Yu et al. (2018b,a)), which generates zenith total delay maps (combined wet and hydrostatic delays) from the High Resolution ECMWF weather model at 0.1-degree spatial and 6-hour temporal resolutions. We convert them to slant range delays using the incidence angle for each pixel.

4.3.2. Ionosphere

We follow the method of Yunjun et al. (2022) and extract total electron content (TEC) values from GNSS-based global ionospheric maps (GIMs) produced by the Center for Orbit Determination in Europe (CODE) to infer the ionospheric phase advance for every pixel at each epoch (Figure SB7). We interpolate TEC values with rotation and extract them at the ionospheric piercing points, assumed to be at an altitude of 450 km, with 20 km sampling. We assume that 85% of the GIM TEC is below the satellite (Sentinel-1 orbits at $\sim$ 700 km), as suggested by an analysis using the IRI2016 model (Bilitza et al., 2017). We then calculate the slant range ionospheric phase advance modified by refraction (Lazecký et al., 2023) and upsample it using bilinear interpolation to provide the correction.

4.3.3. Solid Earth Tides

The long-wave signal from the tides due to lunisolar gravitational forces on the solid Earth has a regular periodicity and can be modelled with high accuracy, allowing correction of InSAR results to prevent temporal aliasing (Xu and Sandwell, 2020). We use the standard model, as implemented in the *earthtide* tool of Generic Mapping Tools (GMT) (Wessel et al., 2019), to calculate vectors of tidal displacements, which we transform to slant range direction and use for correction.

A correction could also be made for ocean tide loading (periodic water mass redistribution from Earth-Moon interaction, (DiCaprio and Simons, 2008)). However, this is more important for regions with large tidal ranges, such as NW Europe, and in any case, the impact on mean velocities for long time series averaged over many years, as in this study, is below 1 mm/year (Wu et al., 2024). Therefore, we do not apply the correction here.

4.4. LiCSBAS: time series inversion and corrections

We perform time series inversion of the input networks of unwrapped corrected interferograms with 0.01×0.01 degree resolution using a Small Baselines Subset (SBAS) method that is both accurate and allows for high spatial coverage in regions with reasonable interferometric coherence (Li et al., 2022). Specifically, we use an updated version of the LiCSBAS algorithm (Morishita et al., 2020; Morishita, 2021) that implements a method that retains some data points that were previously masked due to transient noise or incoherence, on the assumption of a constant velocity.

Prior to the time series inversion, we identify and exclude coseismic interferograms over significant earthquakes (Section 4.4.1), introducing data gaps in the input network. Those gaps are spanned during the later inversion on the assumption of a constant velocity. We identify relatively stable reference areas based on spatio-temporal statistics of the interferometric network and generate interferometric phase loop closure triplets $I_{ABC} = I_{AB} + I_{BC} - I_{AC}$ for all possible combinations of interferograms I from acquisition dates A, B, C. We assume that residuals in triplets formed from correctly unwrapped interferograms should be smaller than π. As our interferometric network is highly redundant, we use the phase loop closure information to identify and drop pixels with unwrapping errors (Shen et al., 2024b; Lazecký et al., 2024). We also allow for possible overall shifts of up to ± 2 phase cycles due to an inconsistent reference area. Additionally, we calculate the mean of the absolute wrapped phase loop error per pixel, as an indicator of potential phase bias (Figure SB10). This is later used as one of the quality measures for the masking of noisy pixels.

Using weights derived from individual coherence maps (Morishita et al., 2020), we perform weighted least squares inversion of the interferometric network into temporal increments that, after cumulating in time, form displacement time series. We store the root mean square error (RMSE) of residuals from this inversion and use its average value per interferogram to identify and drop noisy input interferograms from the network for some frames, then repeat the inversion. The pixel-wise residual RMSE is also used as one of the quality measures for masking of noisy pixels.

After masking noisy pixels using the phase bias indicator, RMSE, and several other quality measures (Morishita et al., 2020), we perform spatio-temporal filtering of the cumulative time series to estimate residual atmosphere that is correlated spatially but not temporally on short time scales (Hooper et al., 2007). This is achieved by highpass filtering in time (using a Gaussian window of width of usually 36 or 72 days, which is three times the average interval between epochs) and low-pass spatial filtering (using a Gaussian window of 2 km width). After subtracting the estimated atmosphere, we estimate the mean velocity using unweighted least squares. We use the original unfiltered time series to estimate the velocity standard deviation using bootstrapping (Efron and Tibshirani, 1986), which we further flatten based on a semivariogram analysis to remove the influence of distance from the reference area (Ou et al., 2022). Finally, we estimate and subtract Eurasian plate motion for all frames (Figure SB11), including those outside Eurasia, by calculating velocities of rotation around the Eurasian plate Euler pole (Stephenson et al., 2022). Some of the key processing parameters are listed in Supplementary Information Appendix C per frame, as they vary, particularly in individual masking thresholds.

4.4.1. Earthquakes

InSAR is good at capturing the deformation associated with shallow crustal earthquakes of moderate to large magnitude (Funning and Garcia, 2019). However, the presence of coseismic displacements results in errors when estimating the long-term velocities associated with interseismic strain accumulation from the time series of interferogram displacements. Therefore, a mitigation strategy is employed to remove the primary effect of the earthquake coseismic deformation field itself. In order to solve for coseismic offsets or truncate the time series to mitigate against such effects, we extract a list of significant earthquakes with locations and depths from the USGS catalogue. We limit ourselves to moderately sized earthquakes with magnitudes $M_w > 6.0$ which are also sufficiently shallow (< 35 km) to potentially produce a displacement signal at the surface (29 events, excluding aftershocks, Table SA3 and Figure SB12). We then manually review each event to ascertain the number of frames affected and decide whether to truncate the time series by starting it after the earthquake (if near the beginning of our time series in 2016) or stopping it before the event (if it occurs towards 2024), or, as in most cases, allowing for an offset in the time series to be solved for across the earthquake date (Table SA3). This greatly reduces the impact of the earthquake

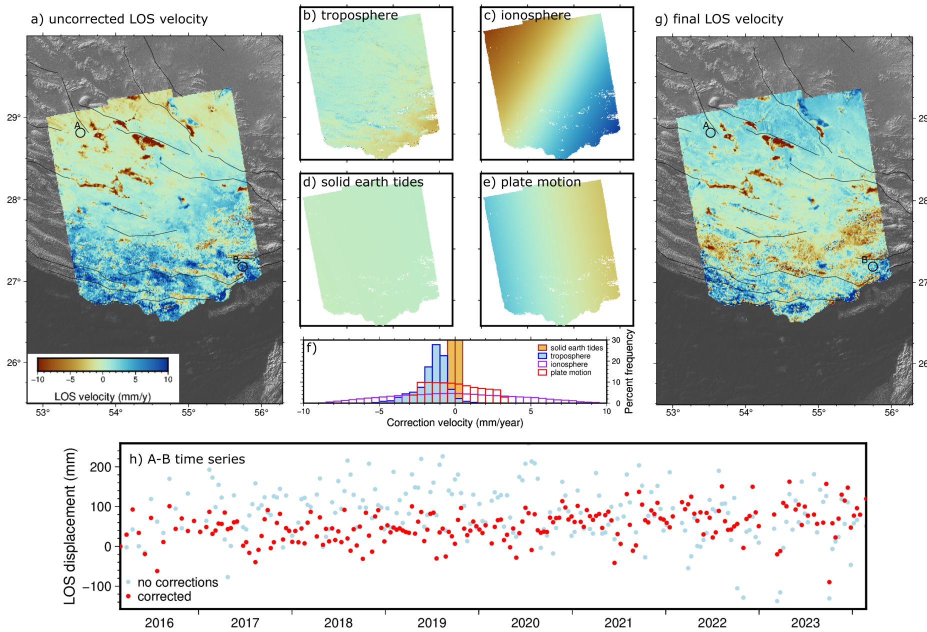

Figure 4: Application of corrections on example frame 130A_06248_212118 over the Zagros mountains of southern Iran: a) LOS average velocity from original time series of uncorrected interferograms; b-d) velocity inverted from time series corrections on troposphere, ionosphere and solid earth tides, respectively; e) plate motion velocity correction (Figure SB11); f) normalized frequency histogram of the b-e correction velocities; g) final LOS velocity with applied corrections; h) relative LOS displacement time series between locations A and B (reference) from both datasets prior and after b-d corrections.

on the velocity estimation, although the impacts of any long-term postseismic signals associated with these events will still be present.

4.5. VELMAP

Following the completion of LiCSAR/LiCSBAS processing, each frame is independently referenced for line-of-sight (LOS) velocity estimation (Figure SB17). To construct a unified velocity field in a Eurasia-fixed reference frame, we integrate these LOS velocities with compiled GNSS data using the approach coded from Wang and Wright (2012) called VELMAP.

We first define the outline of the triangular mesh by applying a 75 km buffer along the outermost frame boundaries and generate the mesh with vertices spaced at approximately 20 km. To enhance computational efficiency and memory management, we divide the mesh covering the AHB into four regions: longitude ranges 0–46°E, 41–75°E, 62–100°E, and 90–120°E (Figure SB13). This segmentation is chosen such that overlapping areas are in regions with good GNSS station density and InSAR coverage. The mismatch between frames in the overlapping regions from separate runs is within 2 mm/yr (Figures SB14, SB15, and SB16).

To constrain the velocity solution in stable Eurasia regions with sparse GNSS data, we introduce synthetic GNSS points at ∼100 km spacing along the mesh edges in the eastern half of our region (Figure 7). These synthetic observations are assigned horizontal velocities appropriate to the plate that they sit on (e.g., $V_e = V_n = 0$ for points on the Eurasian plate), with uncertainties of 1 mm/yr. Vertical velocities for these contrived boundary observations are set to zero with an uncertainty of 3 mm/yr.

The inversion simultaneously estimates velocities at each triangular mesh node, reference frame adjustment parameters, a linear atmospheric correction term based on elevation, and a long-wavelength quadratic ramp for each frame (Wang and Wright, 2012). Three-dimensional (3D) velocities at InSAR and GNSS observation points are interpolated from the triangular mesh nodes. Before the VELMAP inversion, InSAR LOS rates are downsampled to a resolution of ∼5 km. As part of the regularisation, we apply Laplacian smoothing in the inversion, adjusting the relative strength of the smoothing term to achieve a balance between solution roughness and misfit. The combined geodetic solution is then used to generate mosaics of ascending and descending LOS velocities in a Eurasia-fixed reference frame (Figure 6) at the original resolution of 0.01° (∼1 km).

4.6. 3D Velocity and Strain Rate generation

With ascending and descending referenced line-of-sight (LOS) velocities (Figure SB17), we only have two independent components of the 3D velocity field (Wright et al., 2004b). To decompose these into the more readily interpretable eastward and vertical velocities, we follow the approach of Hussain et al. (2018) and Weiss et al. (2020) — we apply a pixel-by-pixel inversion at ∼1 km resolution, with the northward velocity constrained to be equal to the coarse resolution velocity field obtained from the VELMAP inversion (Figure 7a).

We then compute the strain rate fields by taking the gradients of the Gaussian-filtered InSAR eastward velocities (Figure SB20) and the smoothed GNSS northward velocities (Figure 7). The horizontal strain-rate tensor, $\dot{\varepsilon}_h$, represents the symmetric component of the velocity gradient tensor. We derive the second invariant ($\dot{\varepsilon}_{II_h}$), maximum shear ($\dot{\varepsilon}_{shear}$), and dilatation ($\dot{\varepsilon}_{dil}$) of the horizontal strain-rate tensor. These are defined as follows (Savage et al., 2001; Kreemer et al., 2014; Sandwell and Wessel, 2016; Wang and Shen, 2020):

$$\dot{\varepsilon}_h = \begin{bmatrix} \dot{\varepsilon}_{xx} & \dot{\varepsilon}_{xy} \\ \dot{\varepsilon}_{yx} & \dot{\varepsilon}_{yy} \end{bmatrix} = \begin{bmatrix} \frac{\partial V_E}{\partial x} & \frac{1}{2}\left(\frac{\partial V_E}{\partial y} + \frac{\partial V_N}{\partial x}\right) \\ \frac{1}{2}\left(\frac{\partial V_N}{\partial x} + \frac{\partial V_E}{\partial y}\right) & \frac{\partial V_N}{\partial y} \end{bmatrix} \tag{1}$$

$$\dot{\varepsilon}_{II_h} = \sqrt{\dot{\varepsilon}_{xx}^2 + 2\dot{\varepsilon}_{xy}^2 + \dot{\varepsilon}_{yy}^2} \tag{2}$$

$$\dot{\varepsilon}_{shear} = \sqrt{\dot{\varepsilon}_{xy}^2 + \frac{(\dot{\varepsilon}_{xx} - \dot{\varepsilon}_{yy})^2}{4}} \tag{3}$$

$$\dot{\varepsilon}_{dil} = \dot{\varepsilon}_{xx} + \dot{\varepsilon}_{yy} \tag{4}$$

The rotation rate tensor is the antisymmetric part of the velocity gradient tensor. The vorticity vector, ω, is twice the rotation rate tensor (Liu et al., 2018).

$$\omega = \frac{\partial V_N}{\partial x} - \frac{\partial V_E}{\partial y} \tag{5}$$

We have masked out regions where the absolute magnitude of the vertical velocities exceeds 20 mm/yr (Figure 7c), as these values are unlikely to represent tectonic signals. These regions are prevalent across Iran, and there are also large areas in China (east of the Tian Shan and south of Beijing). We have applied the same mask to the eastward velocities to ensure that these anomalous regions do not influence the overall strain analysis. To mitigate short-wavelength noise in the InSAR eastward velocities, a Gaussian filter with a sigma width of 15 pixels (∼15 km) is applied to non-NaN pixels to derive the preferred strain rate fields using a window size of ∼100 km. Although this filtering approach reduces noise, it also damps down any real short-wavelength strain signals visible in the data, for example, due to fault creep (Section 5.3.1).

4.7. LOS Uncertainty Analysis

The mosaics of the line-of-sight (LOS) velocities in the common reference frame, as well as the VELMAP north velocities, allow us to evaluate the uncertainties of our LOS velocities at track overlaps, where we have a redundancy of one LOS velocity measurement ascending or descending (Figure 5). Following Wright et al. (2023), we divided the LOS velocity tracks into four groups, A1, A2, D1, and D2, each containing alternate tracks in the ascending (A) or descending (D) geometry. Instead of using all four groups of LOS velocities and the VELMAP north velocities to derive east and vertical velocities for the entire coverage, we can perform the same operation using three of the groups (e.g., A2, D1 and D2) to produce a version of the east and vertical velocities that partially overlaps with the group left out (e.g., A1). Using east and vertical velocities derived without group A1, we can reconstruct the LOS velocities at pixels overlapping with group A1 using the observation geometries of group A1.

The differences between these two independent estimates of the same LOS velocities allow us to estimate uncertainties of each data set, assuming that the differences resulted from equal uncertainties of the two data sets. The root mean square differences between the observed and reconstructed LOS velocities are 3.7 mm/yr and 3.5–3.6 mm/yr for the ascending and descending geometries, respectively (Figure 5c-f), meaning the average uncertainties are 2.6 mm/yr and 2.5 mm/yr for the ascending and descending LOS velocities.

The slightly lower uncertainty of the descending LOS velocities than that of the ascending LOS velocities is reasonable as the interferograms in the descending tracks are less susceptible to tropospheric and ionospheric noise, being acquired at dawn. In Figure 5, we can see that larger differences tend to occur in areas with snow or vegetation cover, such as the Himalayan front, the Tibetan interior, northern India, northern Tian Shan, and the Caucasus.

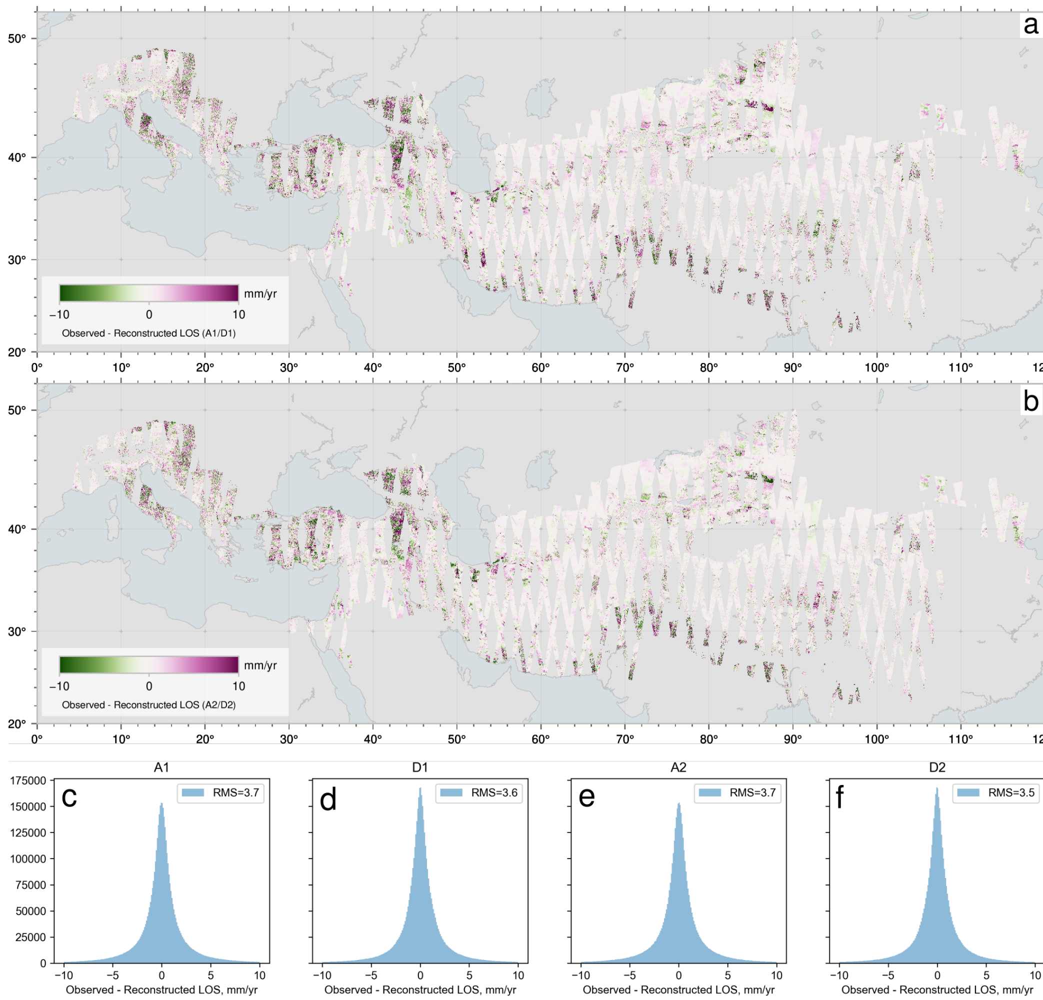

Figure 5: LOS uncertainty analysis through the differences between observed and reconstructed LOS velocities at track overlaps. The observed LOS velocities in Figure 6 are split into four non-overlapping groups of ascending and descending tracks, A1, A2, D1, and D2. (a-b) A1 represents the differences between the observed LOS velocities in A1 and the reconstructed LOS velocities at A1's pixel locations using LOS velocities in A2, D1, and D2. The same applies to D1, A2, and D2, respectively. (c-f) Residual and Root-mean-square (RMS) differences between the corresponding observed and reconstructed LOS velocities.

5. Results

We first describe the line of sight velocity fields that result from combining all the individual frames, before moving on to the 3 components of velocity from the combined decomposition of InSAR and GNSS data. We then show our results for the strain rate calculations for the whole AHB. A zoomed in discussion of the details of the velocity and strain rates for each of the major tectonic region is given in the Supplementary Information. We finish by showing signals associated with creeping faults, postseismic deformation following major earthquakes, and land subsidence due to water extraction.

5.1. Surface Velocities

The Eurasia-fixed ascending and descending LOS velocities (Figure 6) offer a comprehensive view of surface motions across the entire AHB. The most notable feature is the pronounced gradient in velocity across the major fault systems, such as the North and East Anatolian Faults (Figure S1), Main Pamir (Figure S4), Altyn Tagh (Figure S6), Haiyuan, Kunlun, and Xianshuihe Faults (Figure S7). Additional features include localized regions where there is apparent motion away from the satellite in both ascending and descending geometries. This consistent motion in both directions suggests subsidence in these areas, likely due to human activities such as water extraction, as observed in regions such as Türkiye, Iran, and the North China Plain (Section 5.3.3).

The northward velocity field (Figure 7a), largely controlled by the GNSS, reveals north-south shortening as a result of ongoing plate collisions. The overall gradient in the N-S velocity field is approximately 40 mm/yr over a distance of ∼3000 km, with around 20 mm/yr of shortening observed across the Himalayas (Figure 8a). This gradient reflects the intense compressional forces resulting from the ongoing collision between the Indian and Eurasian plates, driving the tectonic deformation in the region. The decomposed E-W velocity field (Figure 7b) reveals large-scale motion across the region, with Anatolia experiencing westward motion at a rate of 20–25 mm/yr relative to Eurasia, and Tibet extending at a similar rate, with rapid rotations around the eastern Himalayan syntaxis. A localized eastward velocity gradient is observed across the major faults, as shown by a ~8000 km profile crossing the North Anatolian, Main Pamir Thrust, Altyn Tagh, and Xiaojiang Fault systems (Figure 8c). The vertical velocity field (Figure 7c) is likely dominated by non-tectonic processes such as groundwater depletion and permafrost and is also noisier and harder to interpret. Values larger than +/-10 mm/yr are unlikely to be tectonically related. The ~4500 km profile (Figure 8e) reveals vertical deformation across the region, with uplift north of the Makran, subsidence around the western syntaxis, Karakoram-Altyn Tagh uplift, subsidence in Qilian Shan due to permafrost, and uplift associated with plateau growth in the north-east.

The comparison between InSAR-derived velocities and GNSS data for the entire AHB shows an RMS of 2.7 mm/yr for eastward velocity (Ve) and 3.5 mm/yr for vertical velocity (Vu) (Figure SB21). For individual regions, the RMS values are 2.9 mm/yr for Ve and 3.8 mm/yr for Vu in Europe (Figure SB24), 3.2 mm/yr for Ve and 4.9 mm/yr for Vu in Anatolia (Figure SB25), 1.7 mm/yr for Ve and 2.8 mm/yr for Vu in the Pamir and Tian Shan (Figure SB26), and 2.6 mm/yr for Ve and 2.5 mm/yr for Vu in Tibet (Figure SB27). These values suggest that, while InSAR captures surface motion with reasonable accuracy, there are still noticeable discrepancies, particularly in the vertical velocity component. The higher RMS value for vertical velocity may be attributed to factors such as atmospheric noise, vegetation, phase bias, or other non-tectonic influences that affect InSAR data. The formal uncertainties from the velocity decomposition of < 1 mm/yr for many areas (Figure SB18), which are propagated from the line-of-sight uncertainties of similar magnitude, do not fully capture the level of noise. The uncertainty estimates from the overlapping frame analysis (Figure 5) of ~3.5 mm/yr and from the longer wavelength referencing from VELMAP (Figure SB19) of 2–3 mm/yr should be considered more realistic.

5.2. Strain Rates

Our strain rate fields of the AHB (Figure 9) shows two styles of deformation, with focused strain concentrated along major fault systems (Figure 9a-b) and more diffuse strain observed away from these large faults (Figure 9c). The major concentrations of strain rate lie along the centre of Greece, North and East Anatolian Faults (Figure S1),

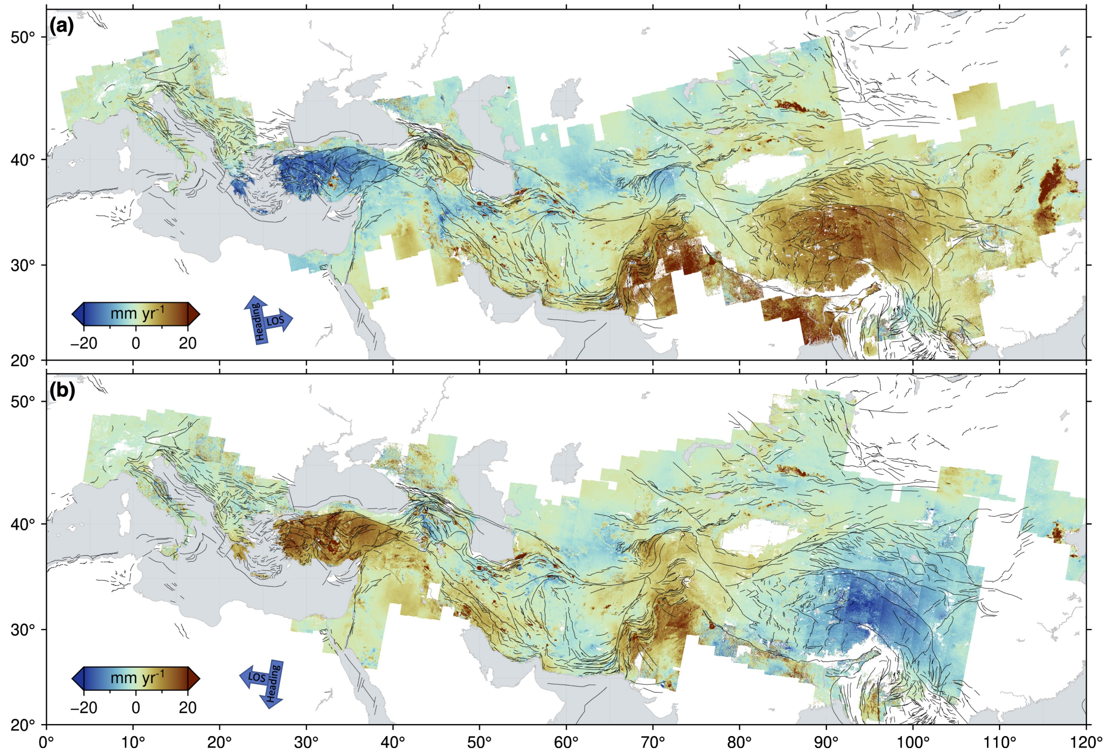

Figure 6: Average mosaic of referenced line of sight velocities for the Alpine Himalayan Belt on (a) ascending and (b) descending orbital heading directions, with positive values indicating motion away from the satellite. Active fault traces (black lines) are from Styron and Pagani (2020).

the Makran and Sulaiman Ranges (Figure S3), the Pamirs (Figure S4), the parts of the Himalayas we can image, and the major strike-slip faults of Tibet (Figures S6 & S7). Distributed strain occurs in regions such as Iran (Figure S2), and the interiors of Anatolia and the Tian Shan (Figure S5) as well as parts of the Tibetan Plateau away from the major faults. Whilst there is noise in these derivative fields, it is possible to discern low straining areas < 20 nstr/yr, in western Europe, Northern Arabia, Central Iran, Afghanistan and much of the region between the Kopeh Dagh of Iran and the Pamirs, as well as North and East of the Tian Shan.

The horizontal dilatation strain rate field (Figure 9c) is particularly noisy, but viewed with a broad brush, positive dilatation, indicating areal increase, supports known extension in Italy, through Greece and Western Türkiye (Figure S1), as well as across the bulk of the southern and central Tibetan Plateau and around the Eastern Syntaxis in Eastern Tibet. Negative dilatation associated with areal decrease and compression is pronounced across the converging regions of the Caucasus, Zagros mountains, Makran, Sulaiman ranges, Pamirs, Tian Shan, Himalayas, and Qilian Shan.

The rotation rates (Figure 9d) reflect the horizontal motion caused by strain accumulation along the major faults, with the direction of rotation depending on the fault's sense of slip (Figure 9d). Left-lateral strike-slip faults are associated with anticlockwise rotation (e.g., East Anatolia, Chaman, Altyn Tagh, Haiyuan, Kunlun, and Xianshuihe Faults), while right-lateral strike-slip faults correspond to clockwise rotation (e.g., North Anatolian and Main Pamir Thrust Faults). These are likely not long-term rotations but reflect elastic interseismic strain accumulation around these faults. The vorticity also captures the long-term clockwise motion of Greece and south-east Tibet around the Eastern Syntaxis. The overall anticlockwise rotation of Anatolia, Northern Arabia, and the Sulaiman Ranges to Hindu Kush is also shown by the broad areas of positive rates of vorticity.

Comparison between GNSS-derived strain rate fields and those from the combined InSAR and GNSS solution shows that InSAR adds finer-scale details, typically localizing more strain onto the large fault systems such as the North Anatolian Fault and those across the Tibetan Plateau (Figures SB29 and SB30). However, in regions of low strain ($<$ 20 nstr/yr), the InSAR can add noise, particularly where coherence is poor. The InSAR strains also contain the contributions from postseismic deformation following past major earthquakes that mostly predate our time series. This postseismic impact is most clearly picked out in the difference of the two maximum strain rate fields (Figure SB29) around Kokoxili on the Tibetan Plateau and the Baluchistan earthquake in the Eastern Makran.

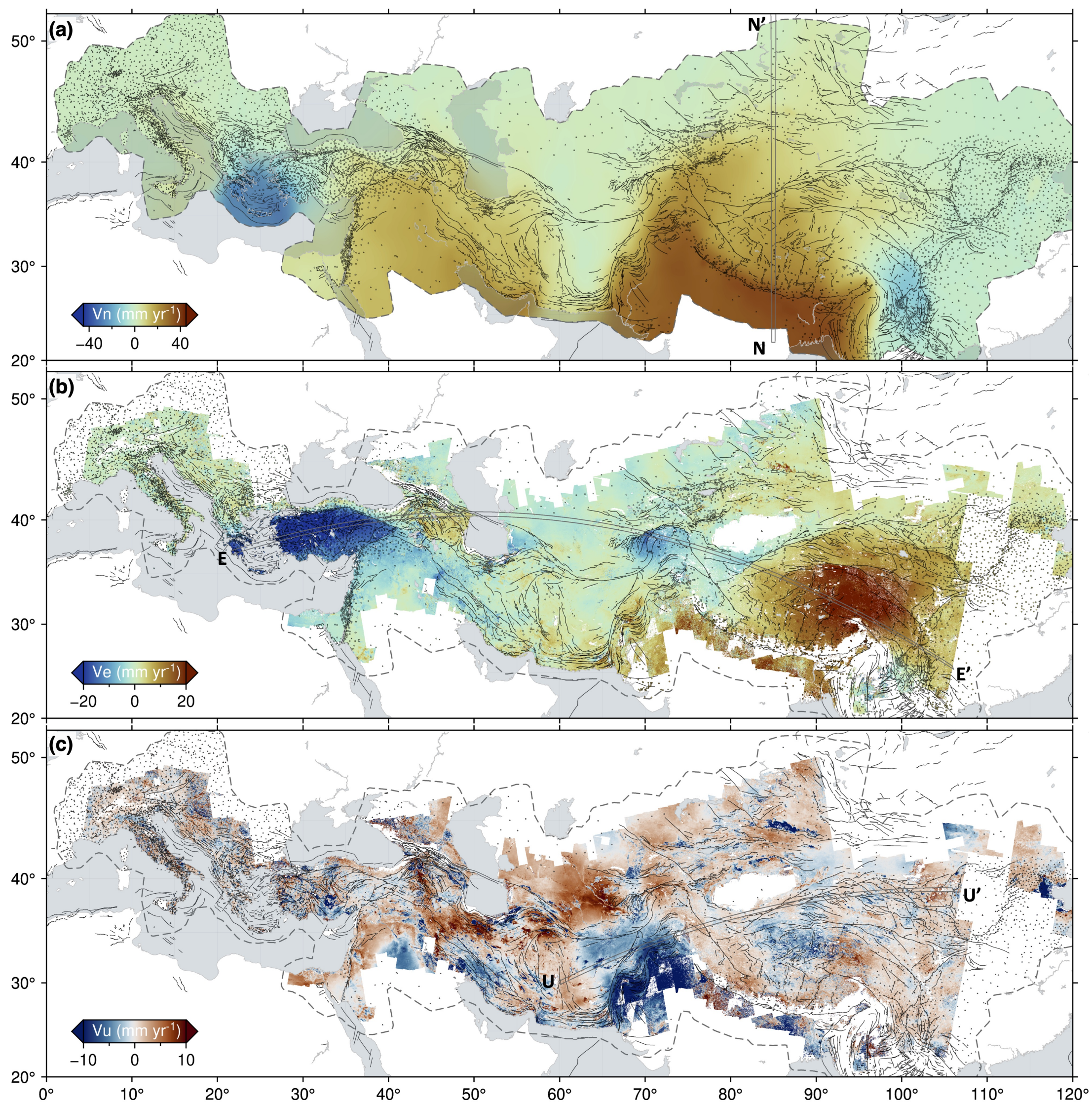

Figure 7: Three components of surface velocity: (a) North, (b) East, and (c) Vertical. GNSS sites are denoted by circles for 2D velocities, triangles for 3D velocities, and squares for locations of pinning synthetic sites along the boundary. The dashed outline shows the extent of the mesh used to align the individual line-of-sight velocities. Gray polygons indicate the extent of the profiles (20 km bins) presented in Figure 8.

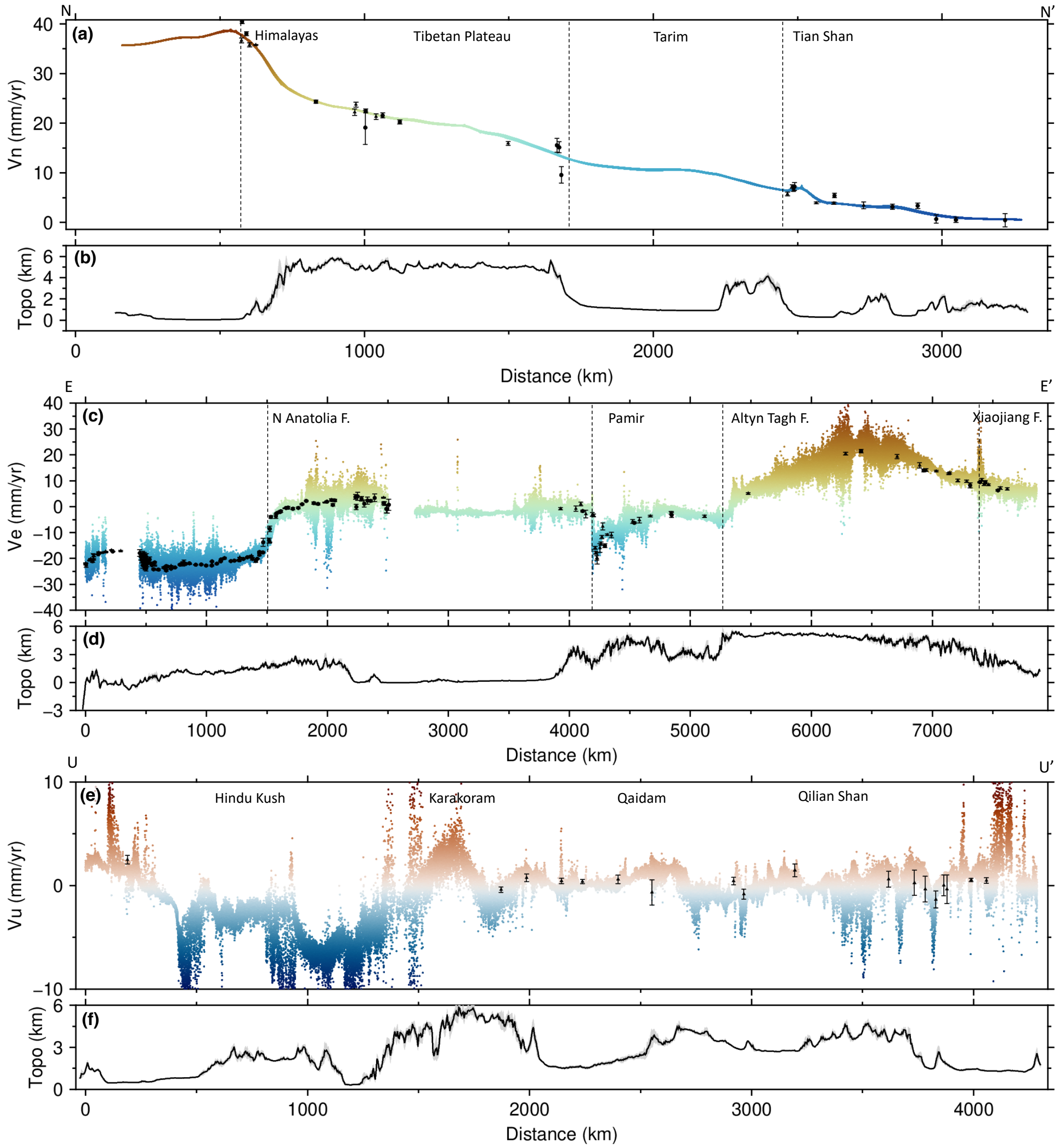

Figure 8: Profiles across the velocity fields of the Alpine Himalayan Belt. (a) Northward component of velocity along longitudinal line 85°E, with GNSS data and topography marked on. (b) Eastward component of velocity from Greece to southeastern Tibet along the Great Circle almost 8,000 km long (20% of the Earth's circumference). (c) Vertical component through Afghanistan to northeastern Tibet. The locations of the profiles are shown in Figure 7. InSAR data are binned into 20 km wide profiles, while GNSS data are binned into 40 km wide swaths.

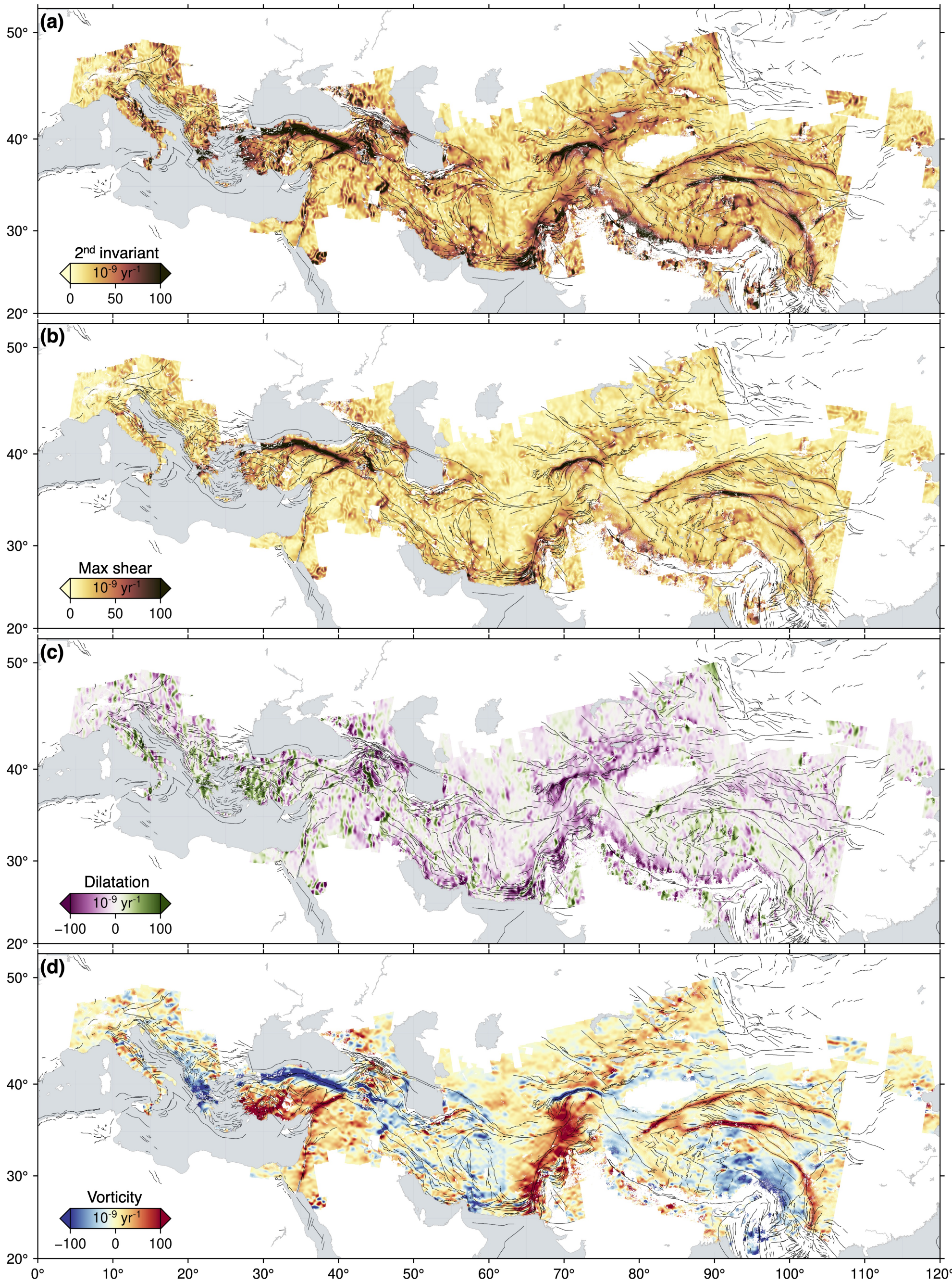

Figure 9: Measures of horizontal strain rates for the Alpine Himalayan Belt. (a) Second invariant of strain rate (nstr/yr), (b) Maximum shear strain rate (nstr/yr), (c) Horizontal Dilatation rate (nstr/yr), where positive indicates extension and negative contraction, and (d) Vorticity (nrad/yr), with positive values indicating anti-clockwise rotation.

5.3. Further Deformation

In addition to the large-scale horizontal velocities associated with long-wavelength continental deformation and the strain of interseismic accumulation across major faults, we also see sharp velocity gradients and high strains associated with individual creeping segments of strike-slip faults and also the elevated strains over regions with past earthquakes. We also observe the largest signals in the vertical associated with non-tectonic signals due to water subsidence across the belt and show a number of examples of these, as well as the influence of permafrost on surface velocities across the Tibetan Plateau.

5.3.1. Creeping Faults

Earthquakes are not the only way for the crust to release its elastic strain. Many previous studies (Avouac, 2015; Harris, 2017; Jolivet and Frank, 2020) have shown that the elastic strain accumulated in the shallow crust can be released aseismically without producing huge seismic waves nor causing sudden movement of the shallow crust. Compared to most active faults, which are fully locked from the surface down to the seismogenic depth during the interseismic period (Reid, 1911), creeping faults are those that are not fully locked; instead, slow, aseismic slip occurs in the periods between earthquakes. Creeping faults sometimes cannot be well identified using GNSS datasets as the GNSS velocities are normally too sparse to distinguish shallow creeping and shallow locking depth and are usually best captured with creepmeters (Bilham et al., 2016). With the help of high-resolution InSAR velocities, we can better capture the near-fault deformation pattern over wider areas, so that we can identify whether a fault is creeping. Across the Alpine Himalayan Belt, previous InSAR studies have identified several creeping faults, which include parts of the North Anatolian fault (e.g., Cakir et al., 2005; Hussain et al., 2016; Jolivet et al., 2023), the Chaman fault (e.g., Fattahi and Amelung, 2016; Barnhart, 2017; Dalaison et al., 2021), the Haiyuan fault (e.g., Jolivet et al., 2013), the Xianshuihe fault (e.g., Qiao and Zhou, 2021) and the Gozha Co fault (e.g., Huang et al., 2023).

By studying our velocity field and strain rates around some of these creeping faults, we observed obvious steps in our east velocity field and high shear strain rates running along them (Figure 10). The pattern of velocity steps and maximum shear strain rates indicates the spatial extent, creeping rates, and creeping depth. Taking velocity profiles perpendicular to the creeping section (Figure 10c,g,k,o), we can better understand the creeping rates and whether the fault is still partly locked (e.g., Hussain et al., 2016; Qiao and Zhou, 2021). The velocity profile across the North Anatolian creeping section shows that the tectonic loading rate (or the slip rate of the fault below the locking depth) on the fault is about 20 mm/yr, which is consistent with previous studies on this section (e.g., Jolivet et al., 2023). The localised offset on the fault (associated with shallow creep) is less than the tectonic loading rate, suggesting the fault is still partly locked, which is supported by previous studies (e.g., Hussain et al., 2016; Jolivet et al., 2023).

Compared with the North Anatolian creeping section, creep on the Chaman fault is much more obvious with a larger spatial extent along the fault (Figure 10e). Our east velocity profile across 31.7°N (Figure 10g) shows that the east component of the tectonic loading rate is about 3 mm/yr, from which we can estimate a fault-parallel loading rate of $\sim$ 7 mm/yr based on a strike of N25°E, which is consistent with the 6.6 ± 1.2 mm/yr estimated by Fattahi and Amelung (2016) and 7 ± 2 mm/yr estimated by Dalaison et al. (2021).

Compared with other creeping faults, the spatial extent of the creeping section is much smaller on the Haiyuan fault in northeastern Tibet (Figure 10i). As the strike of the Haiyuan fault is nearly west-east, our velocity profile indicates that the overall tectonic rates of the Haiyuan fault should be near 4 mm/yr, which is consistent with previous studies (Huang et al., 2022). The offset on the fault suggests a creep rate that is slightly smaller than the loading rate in the location of our profile, suggesting a creep rate of $\sim$ 3 mm/yr, similar to Jolivet et al. (2023).

The Xianshuihe fault also shows a clear step in the east velocity field (Figure 10m) and high maximum shear strain rates between 30°N and 32°N. The east velocity profiles indicate a loading rate of $\sim$ 12 mm/yr, assuming a strike of $\sim$60°, which is close to the 11.5 ± 1.5 mm/yr estimated by Qiao and Zhou (2021). The offset in the east velocity looks comparable with the far-field loading rate in our data.

Our InSAR time series over the Alpine-Himalayan Belt not only provide a way to study the rate of creep but also allow for the investigation of its potential temporal variations. Figure 10q shows a displacement time series for the Chaman Fault for a pixel referenced to another pixel on the other side of the fault (two red points in Figure 10e); because the points are close to the fault, the displacement time series captures how the creep rate is changing temporally. The example time series suggests that creeping on the Chaman Fault is fairly steady in time (~ 3.8 mm/yr), with any variations from steady-state hard to distinguish from noise.

Although a systematic study of creeping faults is beyond the scope of this paper, our results have shown that our derived velocity field and time series can be used for further investigation and comparison of creeping faults within the Alpine-Himalayan Belt. Nevertheless, we can conclude that creep is relatively rare on continental faults.

5.3.2. Postseismic Deformation

During our InSAR data processing, we removed the impact of the coseismic signals that occurred during the acquisition timeframe by adding gaps or truncating the data (Table SA3). However, we did not attempt to correct for postseismic deformation due to earthquakes that occurred before the Sentinel-1 observation period or which occurred within the observation window (although we did truncate time series before some earthquakes that occurred late in the observation window; Table SA3). Some significant postseismic deformation signals are present in our velocity and strain rate data (Figure 11).

Postseismic deformation has been attributed to three primary processes: viscoelastic relaxation, poroelastic rebound, and afterslip (e.g., Ryder et al., 2007, 2011; Wright et al., 2013; McCormack et al., 2020; Zhao et al., 2021). Mapping how deformation varies in space and time following large earthquakes gives an opportunity for understanding the relative contributions of these mechanisms, and can act as a probe to determine lithospheric rheology and investigate the friction properties of faults (e.g., Barbot et al., 2009; Wright, 2016; Ingleby and Wright, 2017; Weiss et al., 2019). Ultimately, understanding postseismic deformation is required if we are to use short-term geodetic data to say anything meaningful about long-term hazard. Due to the variation in earthquake size, earthquake type, and structure of the Earth, the surface deformation pattern of the postseismic phase and its duration can vary significantly between events, although the largest velocities typically decay following a $1/t$ law (Ingleby and Wright, 2017).

Figure 11 shows several examples of signals due to postseismic deformation that we observe in our velocity field. From Figure 11a we could see that there is strong and distributive postseismic deformation due to the 2013 M_w 7.7 Balochistan earthquake (Avouac et al., 2014) during our acquisition time (2016–2024). Previous research showed that the earthquake had a large strike-slip component (Avouac et al., 2014), which is consistent with the $\sim$ 20 mm/yr offset we see in the east velocity field (Figure 11g). The high shear strain rate of over 200 nst/yr on the fault (Figure 11h) suggests that the postseismic deformation on this fault is still dominated by strike-slip movement. The postseismic deformation associated with this fault is thought to be controlled mainly by widespread aseismic slip rather than viscoelastic relaxation (Lv et al., 2022), which is consistent with the deformation pattern we observed.

The 1997 M_w 7.5 Manyi earthquake was three decades before this dataset coverage, but the postseismic deformation related to it is still notable in the velocity field (Figure 11b). Previous studies based on the InSAR time series just after the earthquake suggest that both afterslip and viscoelastic relaxation can be plausible mechanisms for the postseismic deformation related to this earthquake (Ryder et al., 2007). Our velocity suggests the postseismic deformation related to this earthquake does not totally vanish. We consider the $\sim$ 7 mm/yr sharp step in the east velocity field (Figure 11i) to include postseismic deformation rather than purely interseismic based on a previous study that shows that the relative motion on this fault is only 3 ± 2 mm/yr before the earthquake (Bell et al., 2011).

Postseismic deformation related to the 2001 M_w 7.8 Kokoxili Earthquake is also obvious in our velocity field (Figure 11c). This earthquake occurred on the Kunlun fault, and its postseismic deformation generates high maximum shear strain rates of over 150 nst/yr along the fault (Figure 11f,l). Previous studies suggest the possible mechanism for the postseismic deformation of this earthquake to be viscoelastic relaxation and afterslip (e.g., Ryder et al., 2011; Wen et al., 2012), but further separation of these two sources is not well achieved. Moreover, the viscoelastic parameters for the deep crust given from previous studies show large variation (Lv and Shao, 2022). This variation not only comes from different models used in these studies, but also lies in the uncertainties and limitations of the observations themselves. Our derived velocity field, strain rates and time series can provide more constraints to the parameters in future studies.

Postseismic deformation of moderate earthquakes can also sometimes cause deformation that is observable in our velocity field, especially in the line-of-sight velocity results. Figure 11m shows the impact of the postseismic deformation of the 2020 M_w 6.3 Nima Earthquake (Hong et al., 2023) on our line-of-sight velocities. From the time series, the temporal decay in postseismic velocity is clear (Figure 11n).

We chose not to attempt to correct for postseismic signals to obtain "steady-state" velocities, as is often done during GNSS processing (Herring et al., 2016). However, a challenge is knowing what postseismic mechanism to correct for and which earthquakes to include, as we clearly document deformation from events such as the Manyi earthquake, which occurred 30 years prior to our data coverage (Figure 11i). By not correcting for the postseismic signals we see in our data, we provide an opportunity for the community to use the deformation data to build improved models for the events we observe, rather than applying our prejudices on the functional form for the time series corrections.

5.3.3. Non-tectonic vertical signals

In the vertical velocities, we observe non-tectonic deformation with distinct spatial patterns from tectonic deformation. For the most part, these incidences are large areas of rapid subsidence (faster than 10 mm/yr over areas larger than 4 km^2) located in valleys (Gambolati and Teatini, 2015; Herrera-García et al., 2021).

We observe this land surface subsidence in the Po Plain (Bonì et al., 2017), Po Delta (Da Lio and Tosi, 2019), and Tavoliere delle Puglie (Scardino et al., 2022) regions of

Italy; several regions of Western and Central Türkiye (including the Konya Basin and the Küçük Menderes River Basin, Üstün et al., 2015; Yalvaç et al., 2023); Georgia and Azerbaijan; extensively across Iran (Motagh et al., 2008; Goorabi et al., 2020; Mirzadeh et al., 2021; Haghshenas Haghighi and Motagh, 2024); Afghanistan (Meldebekova et al., 2020; Kakar et al., 2025); Pakistan (Khan et al., 2022; Ahmad et al., 2019); and in the Junggar Basin (Wang et al., 2022a) and Beijing regions of China (Zhu et al., 2015; Hu et al., 2019; Chen et al., 2016; Ao et al., 2024).

Previous work has observed the correlation of subsidence rates in the Alpine Himalayan Belt with groundwater depletion, particularly across Iran (Motagh et al., 2008); Kabul, Afghanistan (Meldebekova et al., 2020; Kakar et al., 2025); and Beijing (Chen et al., 2016). However, more recent work derives a more complex picture of land subsidence across, for example, China (Ao et al., 2024), where the mass of cities, groundwater depletion, and hydrocarbon extraction all contribute to rapid land surface subsidence.

In this study, we find over ∼227,000 km^2 is subsiding faster than 10 mm/yr in regions larger than 4 km^2. We choose this area threshold to discount isolated, noisy pixels. We choose this velocity threshold to discount apparently slowly subsiding regions which may instead represent non-deforming noise sources including phase bias (e.g. Ansari et al., 2021; Maghsoudi et al., 2022). Including the whole AHB, the largest region of continuous subsidence is in Eastern China, covering ∼51,000 km^2 (Figure 12b). This region stretches from Beijing in the north, south through Tianjin, finishing in Puyang (Figure 12b). This region alone is similar to or larger than the total extent of land subsidence across the whole of Iran (Haghshenas Haghighi and Motagh, 2024; Payne et al., 2024). In northern China, there is a vast region of subsidence in the Junggar basin, north of the Tian Shan, covering ∼13,000 km^2 (Figure 12a). This figure is similar to the 16,146 km^2 subsidence extent for this region reported by Wang et al. (2022a). Across the whole Tian Shan region, rapid subsidence covers ∼18,000 km^2. Other subsidence regions of significant geographic extent include two regions around the Konya Basin, Türkiye, which total ∼7,000 km^2 in extent (Figure 12c).

We find a total area of ∼1,800 km^2 in which subsidence exceeds 100 mm/yr. The vast majority of these rapidly subsiding regions are located in Iran (1,580 km^2, 88.5 %), but also in Türkiye (123 km^2; near Ödemiş and in the Konya Basin), Pakistan (63 km^2; near Qila Saifullah and Quetta), western Azerbaijan (5 km^2), and in small areas of the Tian Shan (12 km^2).

For analysis of rapid subsidence in Eastern China (longitudes > 111°E), there are limited Sentinel-1 acquisitions in the descending direction, meaning velocity decomposition to retrieve vertical surface motion is not possible through our methodology across large sections of the main subsidence regions here (Figures 7 & 6). Instead, we use only ascending acquisitions and resulting surface velocities in the ascending direction to estimate $\sim$100 km^2 is subsiding faster than 10 mm/yr in Eastern China. We therefore assume for this geographic region that 100% of the recorded ascending direction is vertical. We reproject our GNSS data in this region into the ascending direction to compare to these ascending InSAR velocities (Figure 12b,e). We find reasonable agreement between these velocities ($r^2 = 0.53$), with the strongest disagreement between rapidly subsiding GNSS stations (35-50 mm/yr) and more apparently stable InSAR velocities. This discrepancy may be due to the GNSS instruments that recorded these more rapid velocities being mounted on monuments, for example, buildings, that are subsiding rapidly on a local spatial extent. Conversely, these InSAR results are posted to 1 km pixel widths, thus averaging to some extent deformation in the ascending direction across the pixel. Consequently, rapid subsidence of individual buildings or monuments would not be visible in our InSAR velocities. This discrepancy is additionally highlighted by uncertainties in both GNSS and InSAR velocities. For example, the GNSS site in Figure 12e at around 225 m along the profile has large uncertainties (±10 mm/yr). The spatial pattern of displacement varies significantly within this subsidence region, with the pattern of subsidence appearing like smaller, joined-up subsidence bowls (Figure 12b, e). This pattern suggests that perhaps the depth of aquifer sediments controls subsidence magnitude. Previously, more local work in this region does find that faulting has some control on the subsidence pattern in Beijing (Hu et al., 2019), but future work should investigate these spatial controls on a more regional scale.

Away from Beijing, the largest region of subsidence in the AHB is in the southern Junggar Basin – the Junggar Oasis – in China (Figure 12a). Previous work attributes this subsidence to over-exploitation of groundwater in this arid/semi-arid agricultural region (Wang et al., 2022a). Our results suggest maximum rates of subsidence in this oasis region reach ∼100 mm/yr, comparable to maximum rates estimated by Wang et al. (2022a) using Sentinel-1 over 2015–2020. In terms of spatial structure, the subsidence in this region appears to be bounded along part of its southern margin by the Manasi Fault system (Li et al., 2021b). Similar to the subsidence region south of Beijing, subsidence here is apparent as several smaller, joined-up subsidence bowls, perhaps suggesting a sedimentological rather than structural control on subsidence rate.

Land subsidence in Iran has been investigated thoroughly in previous work (Haghshenas Haghighi and Motagh, 2024; Payne et al., 2024), with the spatial pattern of several subsidence regions controlled by bedrock faults. However, there is a great need to investigate more thoroughly land subsidence drivers, controls on spatial patterns of subsidence in these other critical regions of the Alpine-Himalayan Belt.

On a larger scale, another process that induces land surface subsidence is the thawing of permafrost under a warming climate. Comparing the spatial patterns of our vertical velocities and the model prediction of permafrost presence (Figure 13) reveals a spatial correlation between subsidence and regions with permafrost zonation index (PZI) greater than 0.5, indicating a high likelihood of permafrost presence (Gruber, 2012). In particular, the correlation is best observed in the Qilian Shan and the interior of the Tibetan Plateau, across almost the entire Qiangtang region, consistent with the result of Lemrabet et al. (2023).

At least 5 mm/yr of subsidence is measured across most of the permafrost zone, as averaged over 40 km wide profiles A-A' and B-B'. Up to 10 mm/yr of subsidence rates are measured on the southwest-facing slopes of the Qilian Shan, similar to results obtained by Daout et al. (2021). Up to 20 mm/yr of subsidence is captured within the Kokoxili region in the central Tibetan Plateau, consistent with values of Chen et al.

(2022); Wang et al. (2022c); Lu et al. (2023). This is also where the permafrost zonation index indicates a transition from PZI=1 in northern Qiangtang, where permafrost presence occurs under all conditions, to PZI<0.5 in the southern Tibetan Plateau, where permafrost is only expected under very favorable conditions. We observe this characteristic across the whole region, where permafrost at the border of stability is the most susceptible to thawing and inducing the fastest subsidence. The co-location of sharp transitions in vertical velocity, topography, and permafrost zonation index at ∼340 km on Profile B-B' (Figure 13) is the best testament of this effect.

Our subsidence rates might be susceptible to the potential of underestimation due to unwrapping errors in long-temporal-baseline interferograms from C-band InSAR (Fan et al., 2025) and the potential of overestimation from phase biases due to multi-looking of especially short-temporal-baseline interferograms (Ansari et al., 2021). Despite uncertainties in the absolute subsidence rates measured by InSAR (Liu et al., 2025), we reveal a big picture of subsidence patterns over the most extensive high-altitude permafrost zone on Earth, highlighting the value of InSAR in improving our understanding of high-mountain permafrost dynamics.

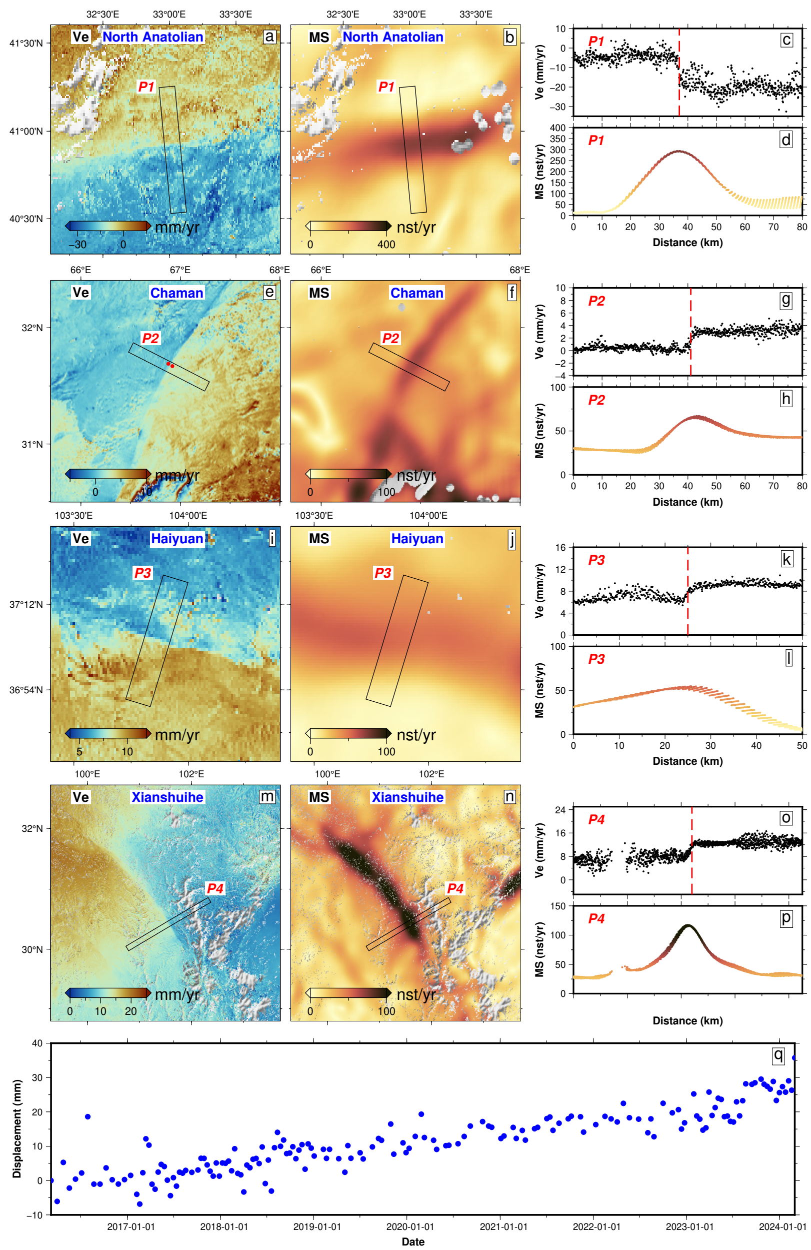

Figure 10: (a,e,i,m) East velocities around the North Anatolian fault, the Chaman fault, the Haiyuan fault and the Xianshuie fault. (b,f,j,n) Maximum shear strain rates around these active faults. (c,d,g,h,k,l,o,p) The velocity and Maximum shear strain rate profiles across these faults. The location of the profiles are shown in the left East velocity and Maximum shear strain rate maps. The width of the profiles is all 10 km. (q) A displacement time series of a pixel referred to a pixel at the other side of the fault based on LiCS frame 042A_05933_131313, and the location of these two pixels are shown by red points in (e).

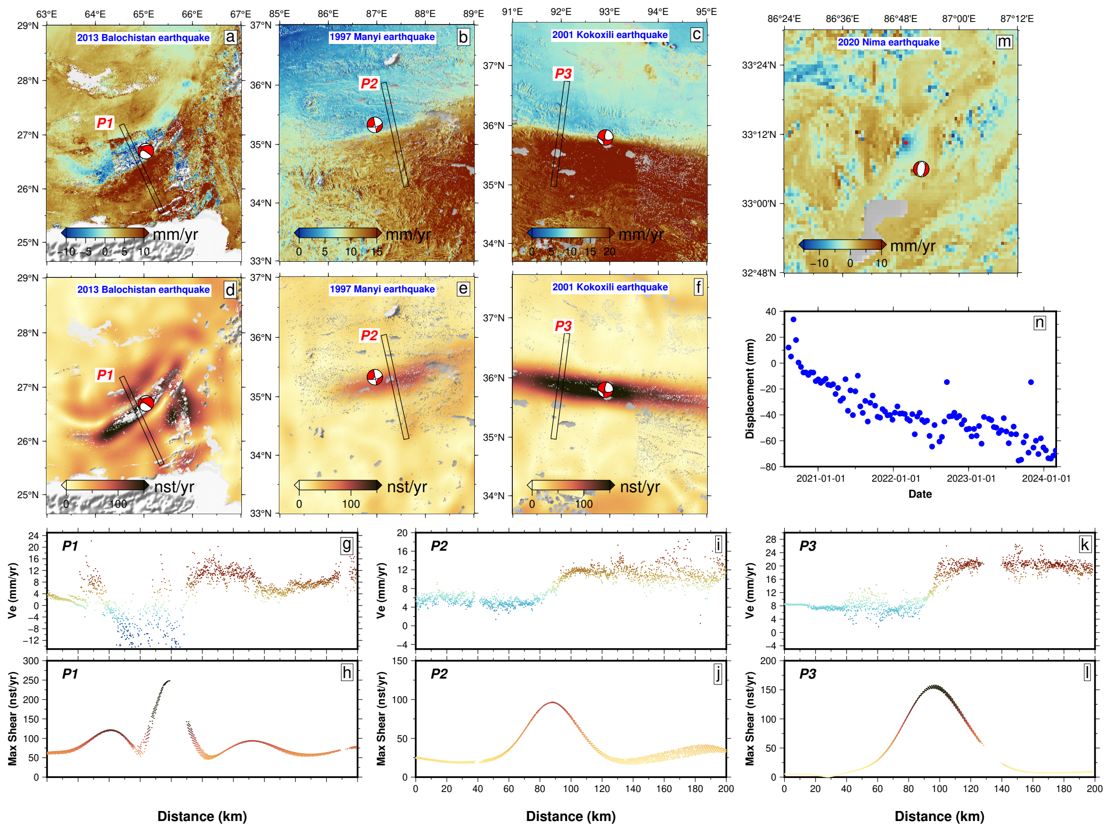

Figure 11: (a,b,c) East velocities around the epicenter of the 2013 M_w 7.7 Balochistan earthquake, the 1997 M_w 7.5 Manyi Earthquake and the 2001 M_w 7.8 Kokoxili Earthquake. (d,e,f) Maximum shear strain rates around the epicenter of these earthquakes. (d-l) East velocity and maximum shear strain rate profiles across the postseismic deformation zones. The locations of the profiles are shown in (a-f). (m) LOS velocities around the 2020 Nima earthquake in the LiCS frame 012A_05642_131313. (n) The postseismic LOS displacement time series for the 2020 Nima earthquake based on the LiCS frame 012A_05642_131313. The location of the pixel is shown by red point in (m). The red focal mechanisms for the earthquakes in (a-f) and (m) are from Global Centroid Moment Tensor Catalog (Dziewonski et al., 1981; Ekström et al., 2012).

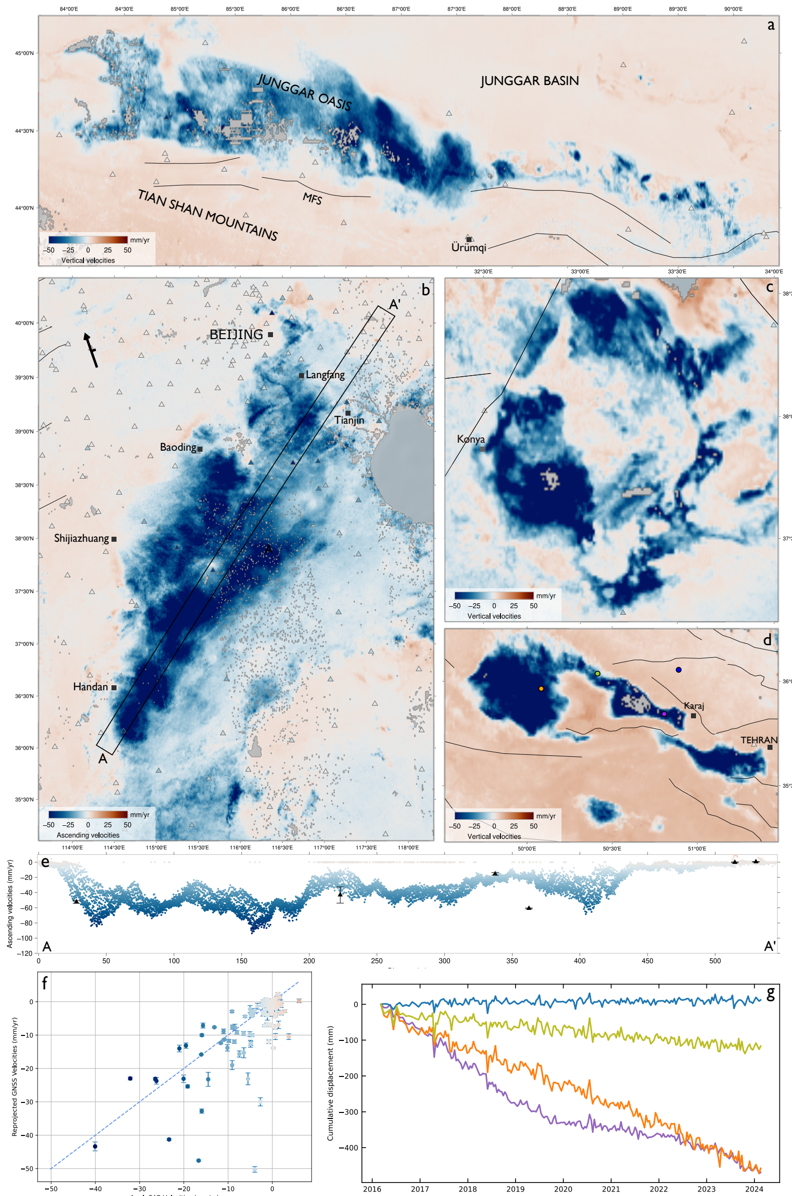

Figure 12: Case examples of rapid vertical subsidence signals associated— according to previous literature or hypothesised association— with groundwater depletion. (a) Tianshan North Slope, China (Yi et al., 2025). (b) The largest continuous subsidence region in this study in Eastern China. Profiled velocities A–A' are in e) (e.g. Ao et al., 2024). (c) Konya Basin, Türkiye (e.g. Üstün et al., 2015)(d) Rapid subsidence near Tehran and Karaj in northern Iran (Haghshenas Haghighi and Motagh, 2024; Payne et al., 2024). (e) Ascending velocities in swath A–A' in b), Eastern China. Triangles = 3D GNSS, circles = 2D GNSS (both this study). (f) Comparison of ascending InSAR velocities from b) at locations of GNSS sites to GNSS velocities reprojected into Sentinel-1 ascending line-of-sight direction. (g) Line-of-sight cumulative displacement time series in the descending direction for frame 035D_05397_131013 at locations in d). Faults from (Styron and Pagani, 2020).

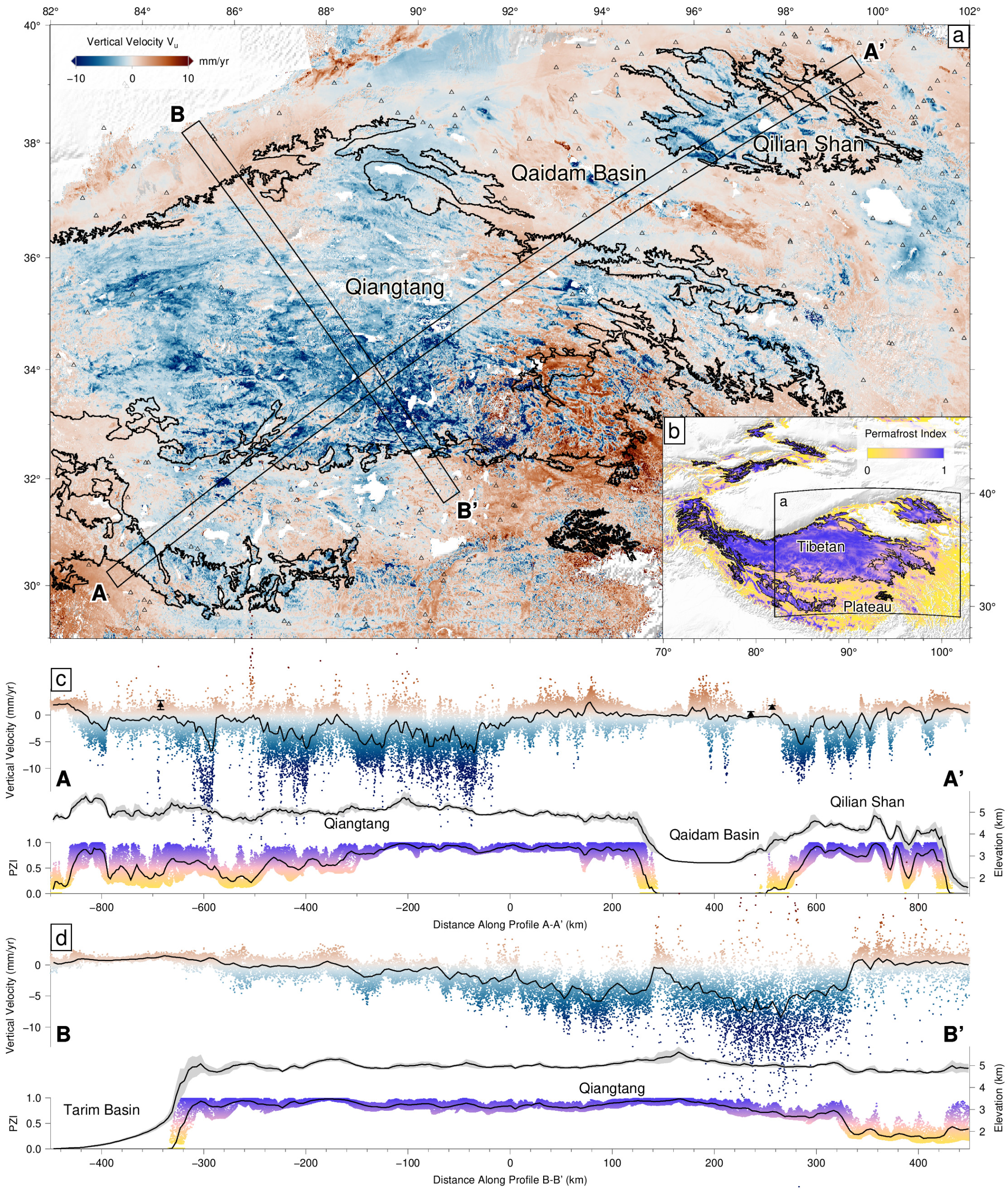

Figure 13: (a) Vertical velocity with 3D GNSS stations shown in triangles and contour line representing Permafrost Zonation Index of 0.5, above which there is a high likelihood of permafrost presence (Gruber, 2012). (b) Permafrost Zonation Index from Gruber (2012). (c,d) Profiles of vertical velocity, the Permafrost Zonation Index and topography.

6. Discussion

6.1. *Validation of Velocity & Strain Rate Fields*

The European Ground Motion Service (EGMS) provides detailed geodetic measurements across Europe and surrounding regions at a resolution of 100 m derived by tying PS InSAR results to a GNSS reference frame (Costantini et al., 2021). Both EGMS and our InSAR show similar eastward velocity patterns (Figure SB22), which is not a surprise given that both are tied to GNSS. Most of Europe exhibits stable velocities with deformation rates of less than ±2 mm/yr. However, in regions like Greece and Anatolia, a significant westward motion is observed at a rate of approximately 20 mm/yr. A comparison between EGMS velocities and our GNSS compilation shows an RMS of 1.1 mm/yr for the eastward velocity and 1.8 mm/yr for the vertical velocity (Figure SB23). When comparing our InSAR-derived velocities with GNSS data in the European region, the RMS values are higher, with 2.9 mm/yr for eastward velocity and 3.8 mm/yr for vertical velocity (Figure SB24). Further work on some of our line-of-sight velocities should improve our noise levels, but some of this difference likely also reflects the difference in spatial resolution of the two processing strategies. We also note, however, that the EGMS time series can be misleading in terms of earthquakes – these create displacement steps in the EGMS time series that are then heavily smoothed, creating the dangerous false impression of pre-cursory deformation anomalies.

The Global Strain Rate Model (GSRM v.2.1) by Kreemer et al. (2014) and our GNSS-derived strain rates generally depict the same broad deformation patterns, with strain distributed over larger areas (Figures 9, SB28, SB29, and SB30). In regions with relatively sparse GNSS measurements, such as the Makran and Sulaiman Ranges, Pamirs, and major strike-slip faults of Tibet, the GNSS-derived strain rate models appear patchy and blurred due to limited data resolution, leading to less precise shear strain rate estimations and reduced clarity in deformation patterns. In contrast, our InSAR-resolution velocities provide higher spatial resolution, localizing more strain onto these structures and capturing finer-scale variations. It is also important to note, however, that our InSAR results also include contributions from postseismic deformation following major earthquakes (Section 5.3.2; Figure SB29).

The GNSS-derived dilatation rate field exhibits smoother and more consistent trends. In contrast, extracting a clean dilatation field from InSAR is more challenging, as it often contains short-wavelength noise. Despite these differences, all models capture similar large-scale deformation patterns, including extension in Italy, Greece, western Türkiye, and central Tibet, and contraction across the Caucasus, Zagros Mountains, Makran, Sulaiman Ranges, Pamirs, Tian Shan, Himalayas, and Qilian Shan.

6.2. Sources of Deformation

By observing continental deformation at high spatial resolution at a trans-continental scale, our work reveals important insights into the variety of deformation styles that we see in the continents. Broadly speaking, our work confirms the emerging view that continental deformation occurs in two modes – strain is focused around major fault structures but more distributed elsewhere (e.g., Wang and Barbot, 2023; Fang et al., 2024a,b). We see this most clearly by contrasting the elevated shear strains we observe around major strike-slip faults in Anatolia (Figure S1) and Tibet (Figures S6 & S7) with the more diffuse shear across regions such as the Iranian Plateau (Figure S2). Because we have uniform InSAR coverage across the entire region, this contrast in deformation style is much clearer than in the previous best strain map on this scale, the Global Strain Rate Model (Figure SB28). We see that the locations of past seismicity from the ISC-GEM catalogue of earthquakes (Storchak et al., 2013, 2015; Di Giacomo et al., 2018) aligns well with our strain rate field (Figure 14). Regions of current (120 year) aseismicity, such as northern Afghanistan and Turkmenistan, northern Arabia and north of the Tian Shan align with regions of very low strain rates in our data. Regions of notable high strain without recent seismicity are the northern Chaman Fault Zone up to the Eastern Syntaxis, and the Altyn Tagh Fault in northern Tibet. Zones of more distributed deformation across the Tian Shan and Makran have perhaps less recent seismicity than we might expect.

It is clear that some of the focused strain we see today is long-lived postseismic deformation from past earthquakes (Section 5.3.2), which would need to be taken into

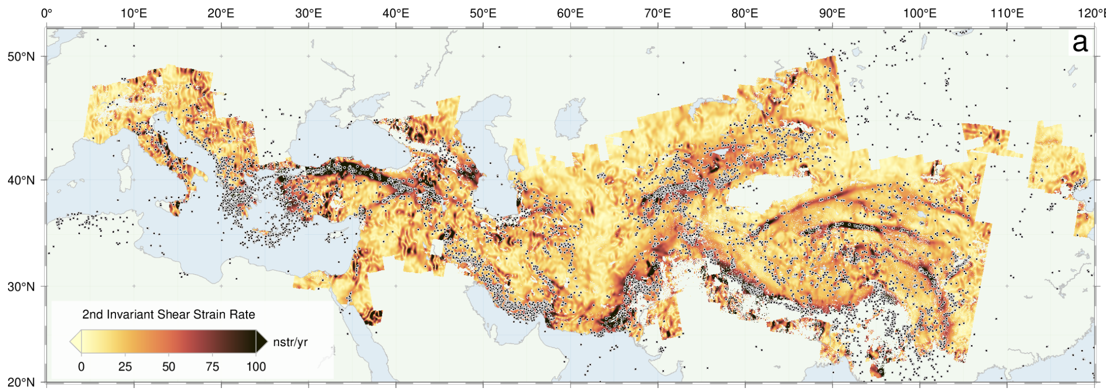

Figure 14: Earthquakes from the ISC-GEM catalogue (Storchak et al., 2013, 2015; Di Giacomo et al., 2018) overlaid on the 2nd Invariant of the strain rate tensor. The earthquakes are from 1904 onwards, and selected for magnitudes (M> 5) at shallow depths (< 35 km).

account in any assessment of interseismic deformation. This transient deformation can last decades - for example, we still see elevated postseismic strain rates on the Manyi Fault in Tibet, which ruptured in 1997. However, it is worth noting that major faults such as the Altyn Tagh Fault, which have experienced long periods without earthquakes, still exhibit significant strain concentrations. Hussain et al. (2018) showed that the strain rate on the North Anatolian Fault is independent of time since the last earthquake, once a decadal postseismic period has ended. Ingleby and Wright (2017) showed that postseismic transients are measurable for up to 100 years, and that the largest postseismic velocities decay following a $1/t$ law (where t is the time since the last earthquake). Our results support the conclusion that, at least for the majority of major faults, rapid postseismic strain-rate transients decay to a steady-state interseismic rate (Meade et al., 2013), rather than varying throughout the earthquake cycle (Wang et al., 2021). One potential exception is the Karakoram Fault in western Tibet – it remains unclear whether the lack of a strain concentration on this fault is due to the low slip rate (Wright et al., 2004a) or because it is late in the earthquake cycle.

The major faults where we see strain concentrations are those that have accrued large geologic displacements. One hypothesis is that repeated earthquakes and significant displacements facilitate strain localisation, through mechanisms such as shear heating, grain size reduction, fabric development, and fluid infiltration. To explain the geodetic observations on major faults requires a weak component of the system (a low viscosity zone or weak fault) capable of producing rapid postseismic transients; this must be embedded within a relatively stronger substrate (relaxation time equal to or larger than the inter-event time) in order to give focused interseismic strain (Yamasaki et al., 2014; Elliott et al., 2016; Wright, 2016). Fang et al. (2024a) estimate a depth-averaged viscosity of 10^{22} Pa s for the lithosphere away from Tibetan faults, which would be consistent with this model.

Accruing significant (10s km) long-term displacements is only possible for long faults that remain in a favourable orientation. They are likely to be weak structures that cross the entire lithosphere (Vauchez et al., 2012). Smaller faults may behave differently, perhaps responding more passively to the overall regional strains, as proposed by Bourne et al. (1998). The interseismic strains associated with slow/small faults are still challenging for satellite geodetic approaches, but we expect future improvements in viewing geometries and coherence (Section 6) to help address whether small faults behave differently to the major structures.

The geodetic data are also capable of observing shallow interseismic creep on some fault structures. This seems to be a relatively rare phenomenon – most faults in the continents are locked and capable of major earthquakes. Even for those faults that do exhibit shallow creep, most are creeping at rates significantly lower than the long-term fault slip rate. Aseismic shallow fault creep is not a significant mechanism for releasing elastic strain accumulation in continental interiors.

Geodetic data are now being increasingly used as inputs to seismic hazard models. One approach is to use the data to help constrain fault-based block models (e.g., Evans, 2022; Styron, 2022). In this approach, the "blocks" each behave as independent microplates; the continuous velocity fields we produce by combining GNSS and InSAR data ensure that there are sufficient data within each block to constrain the free parameters (coordinates of the pole of rotation and rotation rate). A major challenge with this approach is that each block has to be completely isolated by faults, even if such structures do not exist. Alternatively, Johnson et al. (2024) recently proposed a direct strain rate inversion approach; inverting strain rates directly removes the need to isolate "blocks" as the strain rates are independent of rotations. Fang et al. (2024b) showed that to match the observed strain in south-east Tibet required the addition of distributed strain sources, as strain on the mapped faults alone was insufficient. This approach relies on high-resolution strain rate data of the type we provide. Strain rate data can also be used directly to estimate moment deficit rates and hence seismicity rates (e.g., Bird et al., 2010, 2015). Our strain rates are significantly sharper than those derived from the Global Strain Rate Model and we would anticipate that they could be used to make a higher-resolution forecast of seismicity rates.

We note that deformation data are just one component of information required for managing and understanding seismic hazard and risk, alongside seismology, active fault mapping and geomorphology, historical earthquakes, and palaeoseismology (Abdrakhmatov et al., 2024). However, earth observation data and satellite geodesy feed into many aspects of the Disaster Risk Management cycle (Figure SB31). The Continental Strain Rate Fields from InSAR combined with GNSS are now emerging as potentially useful inputs into the preparedness of regions before earthquakes by highlighting the regions straining and providing constraints on models of seismic hazard.

6.3. Challenges & Future Improvements

Working at such a large spatial scale and aiming for millimetre per year accuracy in surface velocities has a number of challenges. The atmospheric noise contributions can be modelled and corrected to a significant degree, and mitigated against through time series analysis to better estimate average velocities. However, there are still some residual effects that could be improved by better and/or higher resolution tropospheric corrections (e.g., Yun et al., 2015), as the tropospheric correction does not always improve the noise in individual interferograms.

The use of InSAR to estimate velocities does introduce finer resolution than that derived from the GNSS in most regions, but it does also introduce noise at short wavelengths in some locations. The InSAR data add the most information in areas which are highly coherent and where there is little existing GNSS coverage, such as across Iran and Central Asia, and within the Tibetan plateau. GNSS site velocities are typically more representative of the displacement of the earth's surface as they are often anchored in bedrock; InSAR is susceptible to the effects of vegetation, loss of coherence (Figure SB9) and phase biases (Ansari et al., 2021), as well as measuring other at, or near, surface displacements not representative of the deeper crustal tectonic motions. Identification of more representative pixels in the InSAR-derived velocity fields will help reduce the noise (as well as using longer time series to better estimate average velocities). Also, it may be possible to roll out corrections for the phase bias from using short period interferograms from individual frames to the whole AHB scale (Maghsoudi et al., 2022). In most cases, however, non-tectonic deformation and phase bias map to the vertical component; horizontal velocities at large scale can only be caused by tectonics.

The short wavelength variability in the surface velocities makes estimation of strain rates challenging. As strain rates are gradients in the velocity, this exacerbates any noise present. This means we have necessarily smoothed our velocities prior to strain-rate calculation to average out some of this dispersion. Future methods could better adapt the filtering approach to account for levels of noise and thus capture the sharpness of velocity gradients expected. In addition, we do not have the north-south component of velocity in the InSAR, relying on the smoother GNSS-derived velocity field. This limitation could be improved in part by using the azimuth (predominantly north-south) motion from burst overlap interferometry (Li et al., 2021a). Of the various strain rates calculated, it has been particularly difficult to capture the dilatation rate sufficiently to make meaningful novel interpretations. This was also found using InSAR applied to the San Andreas Fault (Xu et al., 2021).

Our strain rates show that we are capturing the scars of old earthquakes in the postseismic deformation that is ongoing years and decades after these events. Whilst they are very much part of the velocity and strain rate field, we could choose to account for the postseismic contributions of earthquakes directly in velocity fields, as is done in the GNSS community (Herring et al., 2016; Sobrero et al., 2020), and separate their effects out from the even longer-term deformation. However, doing this in a systematic fashion without biasing the results by our prejudices is challenging – we have chosen to keep postseismic transients in our time series but provide the data that will allow others to make those corrections.

Whilst we are primarily concerned with capturing continental deformation at the scale of the whole AHB, the regions of greatest topography, in particular the Himalayas, are important as locations for major earthquakes. However, it has been very hard to reliably measure the velocities for the highest relief mountains of the Alps, Caucasus, and Himalayas with C-band SAR due to limited coherence at these wavelengths (Figure SB9). This is due to a mixture of environmental impacts including vegetation up to the tree line, and the impact of snow cover at high levels. There is also the challenge of the extreme overall relief, leading to layover and shadowing for radar viewing geometries, which is difficult for unwrapping phase; high relief also results in large gradients in tropospheric delays. It has also been challenging to capture deformation south of the Himalayas in northern India too due to extensive areas of agriculture, and vertical displacement signals due to groundwater extraction.

We expect we can also make general improvements by 1) widening the coverage further across the wider AHB into slower deforming areas, 2) lengthening the time series with more recent data from after March 2024, and 3) producing time series at higher spatial resolution. There is relatively easy potential for 100 m as our catalogue of interferograms has been produced at that resolution, just not the time series and subsequent analysis for this study. By opening up the data set to the community, we give others the opportunity to apply different processing and correction algorithms to our very large observational data set.

6.4. Recommendations

Our results here show that whilst there are concentrations of strain associated with major faults and past earthquakes, an important fraction of the continental deformation occurs over long distances in a smooth, distributed fashion resulting in small strain rates. Ongoing acquisitions to lengthen the Sentinel-1 time series using Sentinel-1C and 1D will be crucial for lowering the threshold for capturing small strain. However, as highlighted with issues of maintaining coherence with C-band, it is important that there are no large data gaps in acquisitions as they break the time series, reducing the accuracy of the average velocities we derive.

Although we have benefited from an exceptionally large data set, some areas lack complete coverage in both ascending and descending viewing geometries, which prevents the retrieval of two components of the velocity field at InSAR resolution. We recommend that acquisitions in both look directions (i.e. including descending) are prioritised over eastern China around the Ordos. Additionally, the region that experienced amongst the largest continental earthquakes in the past century, the M8 Tsetserleg-Bulnay Strike-Slip Earthquake in 1905 in Mongolia (Choi et al., 2018), has a much less complete archive of data, and also should be targeted in the ongoing mission.

The open access data policy from the Copernicus ESA programme for Sentinels has not only been a great benefit in accomplishing a project at this scale, but a necessity. Systematic, regular acquisitions at broad scales are the only way to extract small strain rate signals over long wavelengths typical across the continents, and we recommend that such an approach be continued and adopted by other data providers.

6.5. Outlook for future satellite missions

The current Sentinel-1 mission is set to continue into the next decade (Miranda et al., 2023) with the recent launch of S1C and the launch of S1D scheduled for later in 2025. S1A will be deorbited once S1D becomes operational, but until that time, we anticipate the formation of interferograms between images acquired by S1A and S1C being possible, as has already been demonstrated (Figure SB32). The degradation of the S1A orbit will cause larger baselines between the two satellites than is ideal for deformation measurement, leading to somewhat poorer coherence, but modelling of the degradation indicates that this will not become a significant issue before its end of life, although baselines of 700 m can be expected at higher latitudes (Pinjeiro, 2024).

Over the next decade, several planned ESA missions will lead to improvements in our ability to measure tectonic strain over large areas. The next generation of Sentinel-

1 satellites (Sentinel-1 NG) will also have a repeat interval of 12 days but a wider swath and higher duty cycle, meaning that each patch of ground can be imaged more frequently on average.

The Copernicus expansion mission Rose-L will operate at L-band (Davidson and Furnell, 2021; Petrolati et al., 2023) in the same orbit as Sentinel-1 NG, expanding coherent InSAR coverage in vegetated areas. The longer wavelength will also be advantageous for measuring high-magnitude signals during earthquakes, which may be aliased in Sentinel-1 data.

The joint NASA and ISRO NISAR mission (Kellogg et al., 2020) is due to launch in 2025 and will also acquire L-band with global coverage (Rosen and Kumar, 2021). As well as offering increased coverage in vegetated areas, the left-looking geometry of NISAR will allow for 3-D decomposition when combined with Sentinel-1 data.

The 10th ESA Earth Explorer Mission, Harmony, will consist of two receive-only SAR systems operating in constellation with Sentinel-1D. When combined with the data from Sentinel-1, this will enable the retrieval of a full 3-D deformation field (Prats et al., 2023). As an Earth Explorer mission designed for scientific research rather than an operational mission, it will not be possible to measure 3-D data globally, but Harmony could serve as a proof of concept for future operational missions.

The continuation of Sentinel-1 and future missions will also offer the potential to generate very long time series, by combining data from ERS, ENVISAT, Sentinel-1, Sentinel-1NG, ROSE-L and other missions. This will allow measurement of processes that vary over the long term such as time-varying slip, as was demonstrated for the Haiyuan Fault by combining 20 years of ERS and Envisat data (Jolivet et al., 2013). Ice loss in permafrost across the northeastern Tibetan Plateau was measured over nearly two decades from combining Envisat and Sentinel-1 (Daout et al., 2020). Subsidence and uplift due to water extraction/replenishment in the Beijing Plain was measured spanning three decades from 1992 to 2023 by combining data from ERS, ENVISAT, Radarsat-2 and Sentinel-1 (Fu et al., 2025). The ability to measure changes in the Earth's surface environment over wide areas for decadal timescales will offer new insights into crustal and land surface processes.

7. Conclusion and outlook

We have compiled, aligned and quality controlled 49,608 GNSS velocities collated from 150 recent studies into a Eurasia-fixed reference frame. We have produced the first transcontinental velocity and strain rate field for the Alpine Himalayan belt (20 million km^2), at a spatial sampling of 1 km and with levels of uncertainties in the horizontal $\sim$2–3 mm/yr and we are able to measure strain rates down to $\sim$ 10 nstr/yr, or better in some very coherent regions. We have achieved this by processing large volumes of Sentinel-1 data, applying a number of atmospheric corrections to reduce noise and combining it with the GNSS reference field.

We have found that horizontal velocities are dominated by tectonic processes. Our results show that whilst the horizontal strain rates are concentrated on major strike-slip faults of Anatolia and the Tibetan Plateau and major convergence zones of the Himalayas and Pamirs, the bulk of the Alpine Himalayan orogeny is undergoing diffuse deformation at relatively low strain rates. Significant regions of distributed deformation are found across Western Anatolia, Iran, the Makran, the whole of the Tian Shan and all of the Tibetan Plateau. We observe large-scale anti-clockwise rotations of northern Arabia and Anatolia, as well as along the Chaman Fault zone from the Makran through to the Hindu Kush and Eastern Syntaxis. We observe major broad clockwise rotations around the Western Syntaxis of the Himalayas. Regions of elevated strain are also found to be associated with creeping faults and postseismic deformation from recent major earthquakes. The results agree well with previously derived strain rate fields with GNSS only, but the InSAR provides greater detail on short wavelengths for regions that lack very high densities of GNSS.

We find that the vertical rates are dominated by near-surface processes such as water extraction from aquifers, and permafrost changes at high elevations, except in regions of rapid convergence such as the Pamirs and Himalayas where we observe tectonic uplift.

Improvements we can look forward to include, in the short term, lower uncertainties from increased data with the growing archive with Sentinel-1C and the anticipated Sentinel-1D, and higher-resolution processing better than the 1 km provided here. In the medium term, better coverage from soon to launch L-Band missions, will enable strain to be measured in areas with denser vegetation and higher relief, such as the Himalayas. NISAR's left-looking instrument, combined with the right-looking Sentinel-1, will provide some ability to recover full 3-component velocities.

Finally, one hundred years after Argand (1922)'s observations and tectonic map of the deformation of Eurasia, we look forward to the community developing, testing and applying new/different algorithms to our open access data, and to the resulting improved understanding of continental tectonics.

CRediT authorship contribution statement

Conceptualization: J.R.E., A.J.H. & T.J.W.

Methodology: J.F., M.L. & C.R.

Software: M.L.

Validation: J.F., M.L., C.R., Y.M., D.W. & J.R.E

Formal Analysis: J.F., C.R., Y.M., M.L., D.W., Q.O. & J.P.

Investigation: J.F., C.R., Y.M., M.L., Q.O., J.P. & D.W.

Resources: T.J.W., A.J.H & J.R.E.

Data Curation: M.L., C.R., Y.M., J.F., J.P., Q.O., D.W. & J.R.E.

Writing - Original Draft: J.R.E., J.F., C.R., M.L., D.W., Q.O., J.P. & Y.M.

Writing - Review & Editing: J.R.E., T.J.W. & A.J.H.

Visualization: J.F., J.R.E., D.W., C.R., M.L., J.P, Q.O & Y.M.

Supervision: J.R.E., A.J.H. & T.J.W.

Project Administration: J.R.E., A.J.H. & T.J.W.

Funding Acquisition: J.R.E., A.J.H. & T.J.W.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

We thank the European Space Agency (ESA) for the open data policy on the Sentinel-1 mission. The InSAR data are copyrighted by ESA and are provided through the Copernicus Open Access Hub (https://scihub.copernicus.eu/). The Sentinel-1 InSAR data are also freely distributed by the Alaska Satellite Facility (https://asf.alaska.edu/). The GNSS compilation of 2D and 3D data in Eurasia and ITRF14 reference frames for this study are available here: https://github.com/earjcr1/AHB_GPS. The software *VELMAP* for referencing together InSAR line of sight data with GNSS is available to download here: https://github.com/nerc-comet/velmap. The InSAR data were processed using LiCSAR System, time series inversion was performed using modified LiCSBAS, release version 1.16 from https://github.com/comet-licsar/LiCSBAS. The data files generated here will be made available in a zenodo data repository for download.

Declaration of Generative AI and AI-assisted technologies in the writing process

During the preparation of this work, the authors used ChatGPT for checking of grammar and for coding some algorithms. After using this tool, we reviewed and edited the content and take full responsibility for the content of the publication.

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Supplementary Information — Deformation, Strains and Velocities for the Alpine Himalayan Belt from trans-continental Sentinel-1 InSAR & GNSS

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1. Tectonic Background

The Alpine Himalayan orogenic belt formed from the ongoing collision of Eurasia with the Nubian (African) (Jolivet and Faccenna, 2000; Billi et al., 2023; Nocquet, 2012), Arabian (Reilinger and McClusky, 2011; Allen, 2021; Viltres et al., 2022) and Indian plates (Molnar and Stock, 2009; DeMets et al., 2020). These relative plate motions narrowed and then closed the Tethys Ocean through the process of subduction (Storetvedt, 1990), suturing crustal material with the Eurasian continent over the past few tens of millions of years (Yin, 2010; Parsons et al., 2020; Allen, 2021). Sustained plate convergence uplifted large mountain chains (Ding et al., 2022) and created high peaks such as Everest in the Himalayas at 8,849 m and Jengish Chokusu in the Tian Shan at 7,439 m. The consequent thickening of the crust and the creation of topography generates gravitational potential energy that resists the tectonic driving forces and, along-with the reduction in slab-pull forces (Collins, 2003), results in a rapid slowdown of the rate of northward convergence of the plates with Eurasia (England and Molnar, 2022).

The European Alps are a region of shortening and uplift (Sánchez et al., 2018), whilst the Apennine Mountains running down the length of Italy currently experience extension (Daout et al., 2023; Nucci et al., 2025). Whilst east-west extension is also occurring in east Albania, convergence and thrust faulting along the Adriatic occurs in the western part of the country (Copley et al., 2009). The Southwestern Balkans, north of the Aegean, are at the transition from collision to subduction (d'Agostino et al., 2020), whilst across Greece there is a large velocity gradient associated with the Hellenic Arc and extension within Greece (Taymaz et al., 1991; Chousianitis et al., 2024).

Anatolia similarly exhibits extensional tectonics, particularly in the west; the rapid rotation and westward motion of Anatolia with respect to Eurasia is accommodated along the North and East Anatolian Faults (Weiss et al., 2020; Barbot and Weiss, 2021). Whilst the westerly part of the AHB is most strongly influenced by the relative motion of the Nubian Plate, Anatolia's motion is controlled as well by the northward motion of the Arabian Plate and lateral variations in gravitational potential energy (England et al., 2016).

The Turkish Iranian Plateau (Copley and Jackson, 2006) and Caucasus's active tectonics (Sokhadze et al., 2018; Ismail-Zadeh et al., 2020) are also a result of Arabia moving ~20 mm/yr northward, resulting in a high mountain chain running from the Caspian to the Black Sea. The Turkish Iranian Plateau grows laterally by incorporating the western edge of the Zagros Fold and Thrust Belt (Allen et al., 2013), which runs along the southern margin of Iran and the Persian Gulf. The Zagros is a region of active shortening and seismicity, comprising large scale folds that result in characteristic "whale-back" anticlines (Nissen et al., 2011). The active deformation across Iran is relatively diffuse (Khorrami et al., 2019; Watson et al., 2024) with the Alborz mountains lying along the northern edge with the Caspian Sea as an example of an actively uplifting intercontinental orogen (Ballato et al., 2015). At the eastern end of the country, the tectonics changes from right-lateral shear (Walker and Jackson, 2004) to shortening along the Makran, a region of subduction which is currently accumulating strain and has the potential for megathrust earthquakes (Smith et al., 2013; Penney et al., 2017).

To the north of this lies the complex zone of the Hindu Kush, Afghan-Tajik depression and the Pamirs which are shortening and shearing as a result of the northward motion of India (Schurr et al., 2014; Metzger et al., 2021). The Himalayas run along the southern margin of the Tibetan Plateau and are the arcuate interface between India and Eurasia, accommodating about half of the overall ~40 mm/yr shortening between the two (Ader et al., 2012). With the largest earthquake potential and some of the greatest density of human populations, this stretch of the AHB presents some of the biggest earthquake risk (Bilham, 2019).

The Tibetan Plateau has grown to 5 km elevation from the sustained collision, shortening, and thickening of the crust (Yin and Harrison, 2000; Royden et al., 2008), and, apart from thrust faulting in the north-east and east, it is now dominated by strike-slip and extensional faulting within it. Beyond all of these mountain ranges, deeper within Eurasia, the Indian plate's most northerly significant tectonic impact is in shortening occurring within the Tian Shan (Jolivet et al., 2010; Zubovich et al., 2010).

2. Results - Active Tectonic Regions of Interest

We consider each of the main tectonic areas of significant deformation in turn, running from west to east, highlighting the general kinematics we see and the significant localisation of strain that we calculate.

2.1. Türkiye

Türkiye, located at the intersection of the Eurasian, African, and Arabian plates (Figure 1), is characterized by complex and high magnitude tectonic deformation. Anatolia experiences a push from the west, where Arabia is colliding with Eurasia, and a pull from the west, where subduction occurs beneath the Hellenic Arc (Barka and Reilinger, 1997; Taymaz et al., 2007; England et al., 2016). The motion is accommodated in part by motion on two major strike-slip faults - the right-lateral North Anatolian and left-lateral East Anatolian Faults (Şengör et al., 2005; Güvercin et al., 2022), with extensional faulting dominating in western Anatolia (Yilmaz and Gürer, 2024). Although Anatolia has often been described as a microplate (e.g., McKenzie, 1972), geodetic studies from GNSS show non-negligible internal deformation (e.g., Aktug et al., 2009).

Türkiye has experienced a large number of damaging recent and historical earthquakes, which have impacted all the tectonic zones described above. The North Anatolia experienced a westward progression of $M_w > 7$ earthquakes in the 20th century (Barka, 1996), culminating in two large earthquakes in 1999 (e.g., Özalaybey et al., 2002; Bürgmann et al., 2002). In the last few years, the East Anatolian fault experienced several major earthquakes, including the 2020 Mw 6.8 Elazığ earthquake (Tatar et al., 2020), and the two devastating earthquakes (Mw 7.8 and 7.6) of the 6 February 2023 Kahramanmaraş sequence (e.g. Hussain et al., 2023; Ergintav et al., 2024). Western Anatolia has experienced smaller but nevertheless damaging earthquakes, including the normal-faulting 1995 Dinar earthquake (Eyidoğan and Barka, 1996; Wright et al., 1999).

A large number of previous studies have used InSAR and GNSS to study velocity and strain rate in different parts of Türkiye, and we do not attempt to provide a comprehensive review here. We focus on those studies that have attempted to map deformation across the whole of Türkiye. The most recent GNSS compilation from (Kurt et al., 2023) contains velocities from 836 stations. The InSAR velocity field from Weiss et al. (2020) uses the first ~5 years of data from Sentinel-1 to derive east-west and vertical velocity fields. Both of these studies capture the large-scale features expected from the tectonics of the region, and Weiss et al. (2020) also shows significant non-tectonic vertical deformation, particularly associated with the Konya basin in Central Anatolia. Other studies have focused on specific aspects of the deformation field, such as the dynamics of the creeping section of the North Anatolian Fault (e.g., Cetin et al., 2014; Jolivet et al., 2023)

Our velocity field (Figure S1) improves on the results of Weiss et al. (2020), using more sophisticated processing and a longer time series. The shear strain rates associated with the North and East Anatolian Faults are sharper, being derived directly from gradients of the referenced InSAR velocities. This is also seen very clearly in the profiles crossing the faults (Figure S1). The cleaner velocity fields show that creep is occurring on parts of the East Anatolian Fault as well as on the well-known sections of the North Anatolian Fault – further work is needed to understand how the creep rate was influenced by the 2020 earthquake.

Our results corroborate previous studies that show all of the North and East Anatolian Faults having focused, elevated strain rates, which appear to be independent of the time since the last earthquake. This suggests that these major faults reach a steady state interseismic strain rate (Hussain et al., 2018). A possible exception is the south-western end of the East Anatolian Fault, which ruptured in the Kahramanmaraş sequence. While there is elevated strain in this region, seen both in the shear strains and rotations (Figure S1), the strain rate is lower than on the north-eastern part of the East Anatolian Fault. This may simply reflect the fact that the fault splits here into multiple strands, so the strain is more distributed.

The horizontal dilatation field shows very clearly the high rates of extension in Western Anatolia, associated with normal faulting there. A long profile through our data crossing all of Anatolia (Figure S1) also shows clearly the steadily increasing eastward velocities as you go from east to west across the region - demonstrating that Anatolia does not behave as a rigid microplate.

While the east-west and north-south velocities are well constrained by both InSAR and GNSS data, the vertical component is less certain due to phase biases and potential land subsidence effects. The 3D GNSS data, which could provide better constraints on the vertical velocity, also show higher uncertainty in these regions, which necessitates a more cautious interpretation. Nevertheless, we see high rates of subsidence in the Konya basin (Figure 12) confirming the result of Weiss et al. (2020).

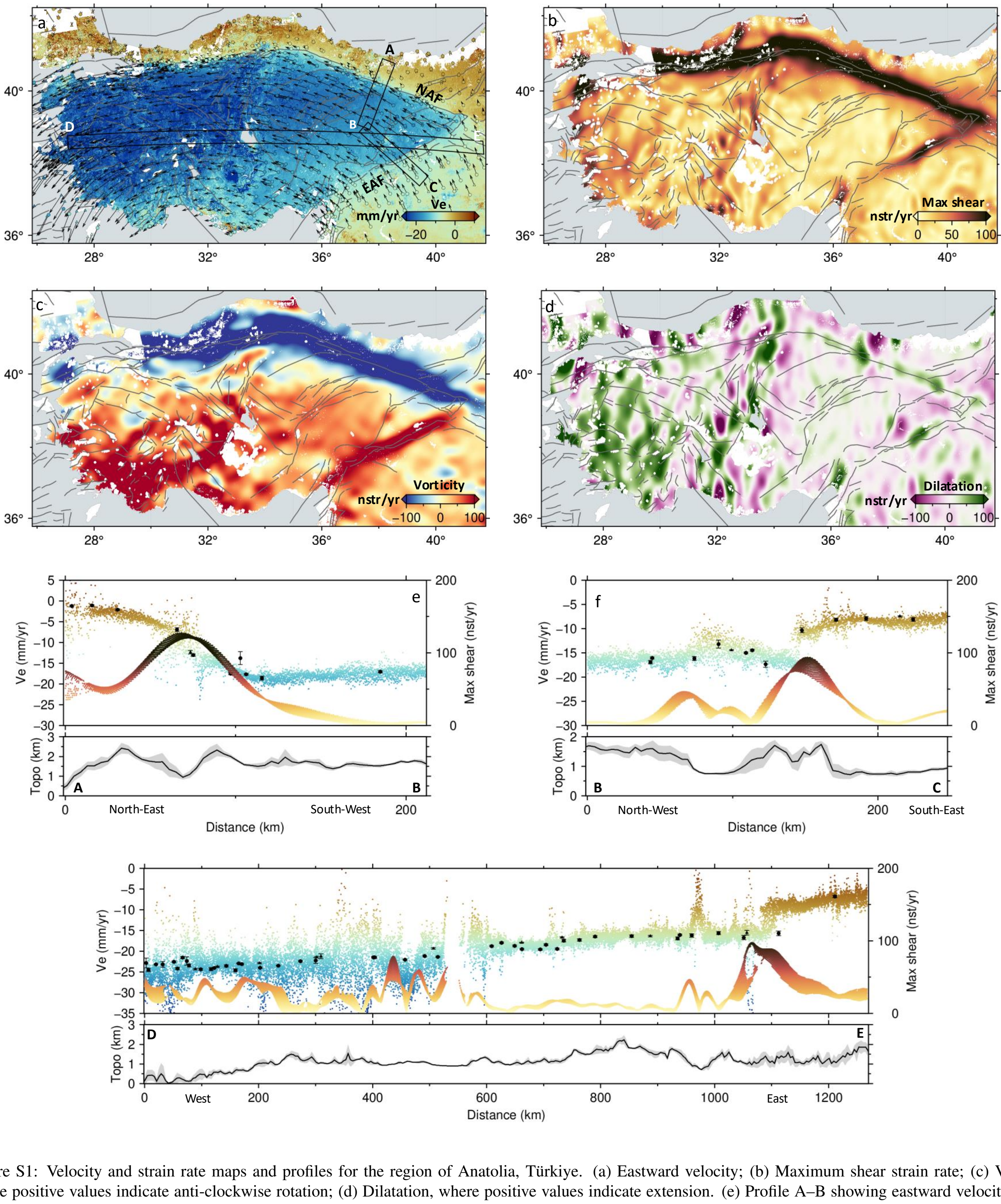

Figure S1: Velocity and strain rate maps and profiles for the region of Anatolia, Türkiye. (a) Eastward velocity; (b) Maximum shear strain rate; (c) Vorticity, where positive values indicate anti-clockwise rotation; (d) Dilatation, where positive values indicate extension. (e) Profile A–B showing eastward velocity across the North Anatolian Fault (NAF) and corresponding maximum shear strain rate; (f) Profile B–C showing eastward velocity across the East Anatolian Fault (EAF) and corresponding maximum shear strain rate; (g) Profile D–E showing eastward velocity from west to east and corresponding maximum shear strain rate.

2.2. *Iran*

Much of Iran's active tectonic deformation is driven by the northward and anti-clockwise motion of the Arabian plate colliding with the south-west of the country. This collision contributes to the active shortening of the Zagros Fold and Thrust Belt in south-west Iran (Nissen et al., 2011), which accommodates about half of Arabia's northward motion (Hollingsworth et al., 2006). Previous research finds deformation in much of the rest of Iran is concentrated in the Alborz and Kopeh Dagh mountains, with Central Iran and the Lut desert behaving near to aseismically (Hollingsworth et al., 2006).

Overall, we find, in agreement with previous studies, that active deformation and subsequent strain are relatively diffuse across Iran (Khorrami et al., 2019; Watson et al., 2024). Where there is elevated shear strain (Figure S2b, c), this strain forms wide bands and is apparently not tightly localised to single faults but wider deforming fault belts. This is particularly true in the Zagros belt. Strain in this fold-thrust belt does not exceed ~50 nanostrain/year and is spread diffusely across much of the entire width of the Zagros (~300 km^2).

Stronger strain localisation is evident in north-east Iran on the margins of the Turan platform. Our results indicate this platform is moving to the west, with the rate of westward motion increasing from east to west to a maximum of ~8 mm/yr. This westwards motion is accommodated by the north-west south-east striking Main Kopeh Dagh and Ashkabad right lateral strike-slip faults on the platform's northern extent (Hollingsworth et al., 2006; Walker et al., 2021), and a series of smaller faults to the south, including the Robet-e Qarabil Fault (Figure S2). Strain on these southern faults reaches ~70 nanostrain/year.

In north-west Iran, we observe right-lateral strike slip motion across the North Tabriz fault (Figure S2e, Copley and Jackson, 2006). Our results indicate that, in the far field, areas north and south of the Tabriz fault are moving ~3-4 mm/yr, towards the east on the north and to the west on the south. On the fault itself, we find some elevated strain compared to neighbouring regions at ~45 nanostrain/year (Figure S2). Some of this shear strain motion may be post-seismic deformation associated with two series of M_W4–6 earthquakes in February 2020 and late 2023 (near Khovy on the Iran-Türkiye border). Overall, however, the strain and east-west motion across this fault is still quite diffuse (reaching 40 nst/yr over 100 km, Figure S2e) despite the influence of post-seismic signals. This diffuse strain provides evidence for strong strain partitioning in this part of the Alpine-Himalayan Belt (Copley and Jackson, 2006).

Finally, we observe significant post-seismic deformation in north-west Iran associated with the November 2017 Sarpol-e Zahab earthquake in the north-west Zagros mountains (Figure S2e, 300 m). We have removed all co-seismic interferograms associated with this earthquake, so any residual deformation around this rupture site is due to post-seismic relaxation. Overall, the wider region of this earthquake is moving ~4-5 mm/yr to the west, meaning post-seismic velocities in the east component direction since November 2017 are ~10 mm/yr for both post-seismic lobes. Previous work has attributed this post-seismic deformation to afterslip (Fathian et al., 2021). We note that this elevated deformation is associated with elevated shear strain around deforming lobes, reaching ~80 nstr/yr at this location.

Overall there is good coverage in our Iran-wide velocity field due to high coherence (Figure SB9), in part due to the lack of extensive vegetation towards inland regions of the country. This lack of vegetation additionally contributes to a low level of noise in the velocities (Figure S2a). This is particularly evident in velocity profiles of desert regions in the north-east of the country at the border with Turkmenistan (Figure S2c) where the spread of values is 2 mm/yr and the retrieved strain rates are very low (< 10 nstr/yr).

2.3. *Makran & Hindu Kush*

The Eastern Makran, Sulaiman Ranges, and Hindu Kush are a tectonically complex region at the western edge of the Indian plate (Figure 1). The Chaman fault and associated sub-parallel fault systems runs along much of this boundary accommodating the fast relative motion of India with Eurasia of ~ 30 mm/yr (Szeliga et al., 2012). The Chaman fault is unusual in that it has not experienced any major historical earthquakes, despite being a major strike-slip zone, although parts of it are creeping (Dalaison et al., 2021). The Hindu Kush region represents an unusual tectonic setting characterized by ongoing slab break-off processes, as evidenced by intermediate-depth seismicity. The Indian lithospheric slab is subducting northward beneath the Hindu Kush, with the slab's penetration depth increasing along its strike (Kufner et al., 2021; Metzger et al., 2021). The Sulaiman Ranges in western Pakistan are a fold and thrust belt that experience a style of shallow crustal faulting that suggests it is deforming out over the Indian foreland with shear along its edges (Reynolds et al., 2015). The Makran region is a wide accretionary prism (Smith et al., 2012) that results from the subduction of the Arabian plate beneath eastern Iran and Pakistan that has the potential for major earthquakes along its interface (Penney et al., 2017). At the eastern end of the Makran, the 2013 M_w 7.7 Baluchistan earthquake ruptured the arcuate Hoshab fault (Avouac et al., 2014).

The coherence in this region (Figure SB9) is amongst the best seen across the AHB in the are in the north-west in Afghanistan, and amongst the worst in South-east Pakistan where agriculture around the Indus river is challenging for C-band to maintain coherence. It is also the region with the least number of GNSS sites (Figure SB1), leading to some large estimates of uncertainty in the derived 3-component velocity (Figures SB18 & SB19), but it is where InSAR offers the potential to improve estimates of velocity and strain rates the most (Figure S3.

In our results, the bulk of Afghanistan is not moving east-west with respect to Eurasia, but along the Makran, Sulamain Ranges and Hindu Kush we see eastward velocities of ~ 10 mm/yr. The shorter wavelength change in velocities from east to west in the Eastern Makran is due to postseismic deformation following the 2013 M_w 7.7.Baluchistan earthquake (Lv et al., 2022). The near-zero east velocities at the north-western edge of the area in Figure S3a illustrate the low level of noise possible under the most ideal circumstances, with spreads in the velocities of < 1 mm/yr (Figure S3c). The vertical competent of velocity is much more difficult to reliable interpret, especially with the lack of GNSS constraints (Figures S3d & 7c) which possible leads to some long wavelength errors.

The maximum shear strain rate is dominated by the earthquake postseismic in Baluchistan, but also broadly aligns with the Chaman Fault zone. Due to its strike running almost north-south, we poorly resolve the expected concentration of shear here despite the overall relative motion > 30 mm/yr, as our northward velocities are relatively smooth (Figure 7) have come largely from the very sparse GNSS. Across Afghanistan, we observe very low strain rates < 20 nstr/yr and north of the Herat Fault, we observe near-zero strain in our data indicating the level of our noise floor for this region.

The whole region running along the Chaman Fault Zone experiences a very broad area of anticlockwise rotation, as viewed in the vorticity for the belt (Figure 9), as a result of the northward motion of India.

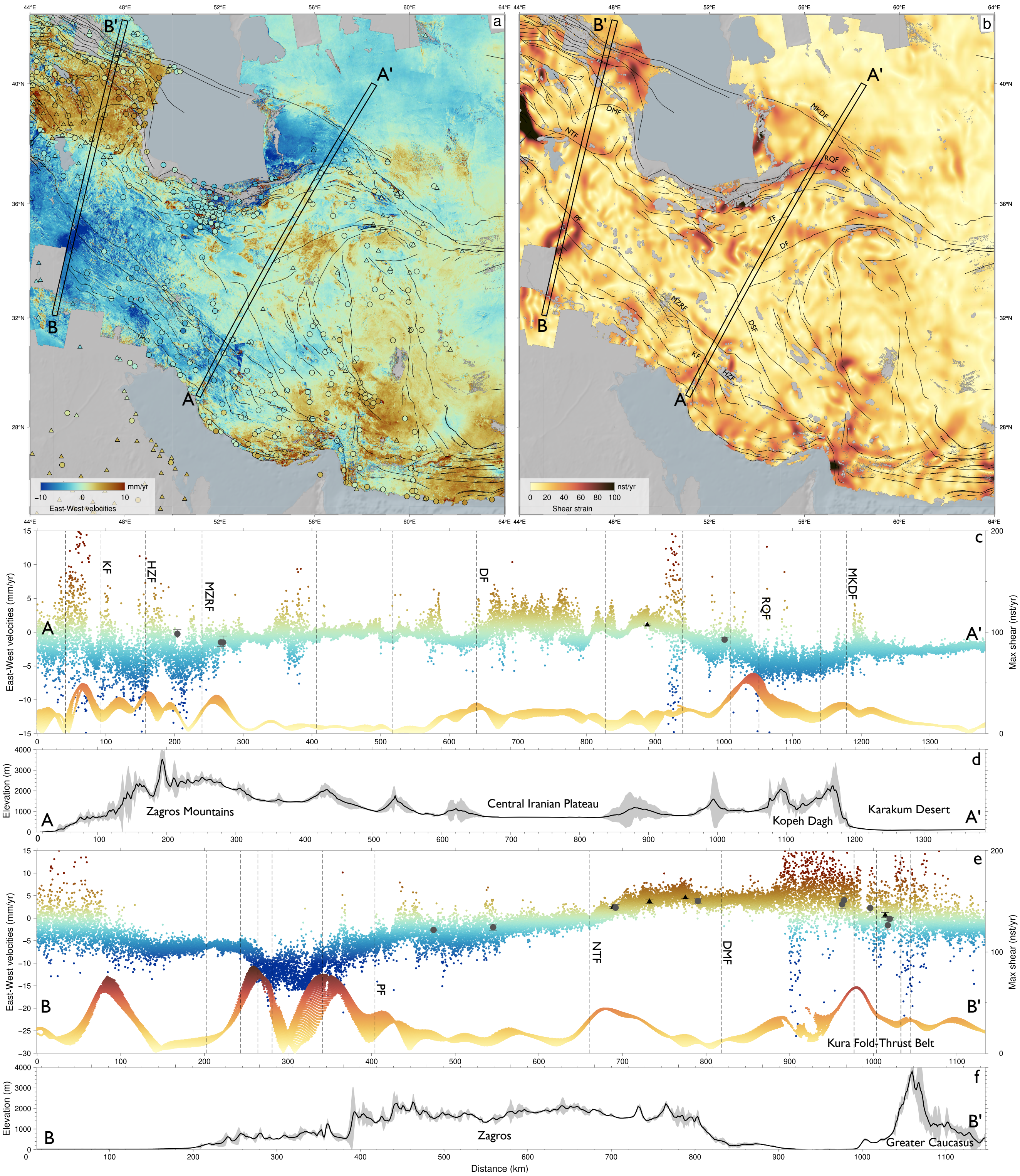

Figure S2: Velocity and strain rate maps and profiles for the region of Iran. (a) Surface velocities of Iran in the East-West direction. Blue, negative velocities indicate motion towards the west; brown to the east (Watson et al., 2024). Circles indicate East-West GNSS velocities compiled in this study. Triangles = 3D velocities; circles = 2D velocities. Fault names from (Hessami and Jamali, 2006); fault locations from (Styron and Pagani, 2020) (KF = Kazerun Fault, HZF = High Zagros Fault, MZRF = Main Zagros Reverse Fault, DSF = Deh Shir Fault, DF = Doruneh Fault, TF = Torud Fault, RQF = Robat-e Qarabil Fault, EF = Esfarayen Fault, MKDF = Main Kopeh Dagh Fault; NTF = North Tabriz Fault; PF = Piranshahr Fault; DMF = Dasht-e Moghan Fault). (b) Maximum shear strain: sig15, window 100. (c) East-West velocities and maximum shear strain in swath A–A' in a) and b). Black dashed lines denote faults mapped in a) and b). (d) Topography in swath A–A' in a) and b). Triangles = 3D velocities; circles = 2D velocities. (e) Profiled East-West velocities and maximum shear strain in B–B' swath. (f) Topography in B–B'.

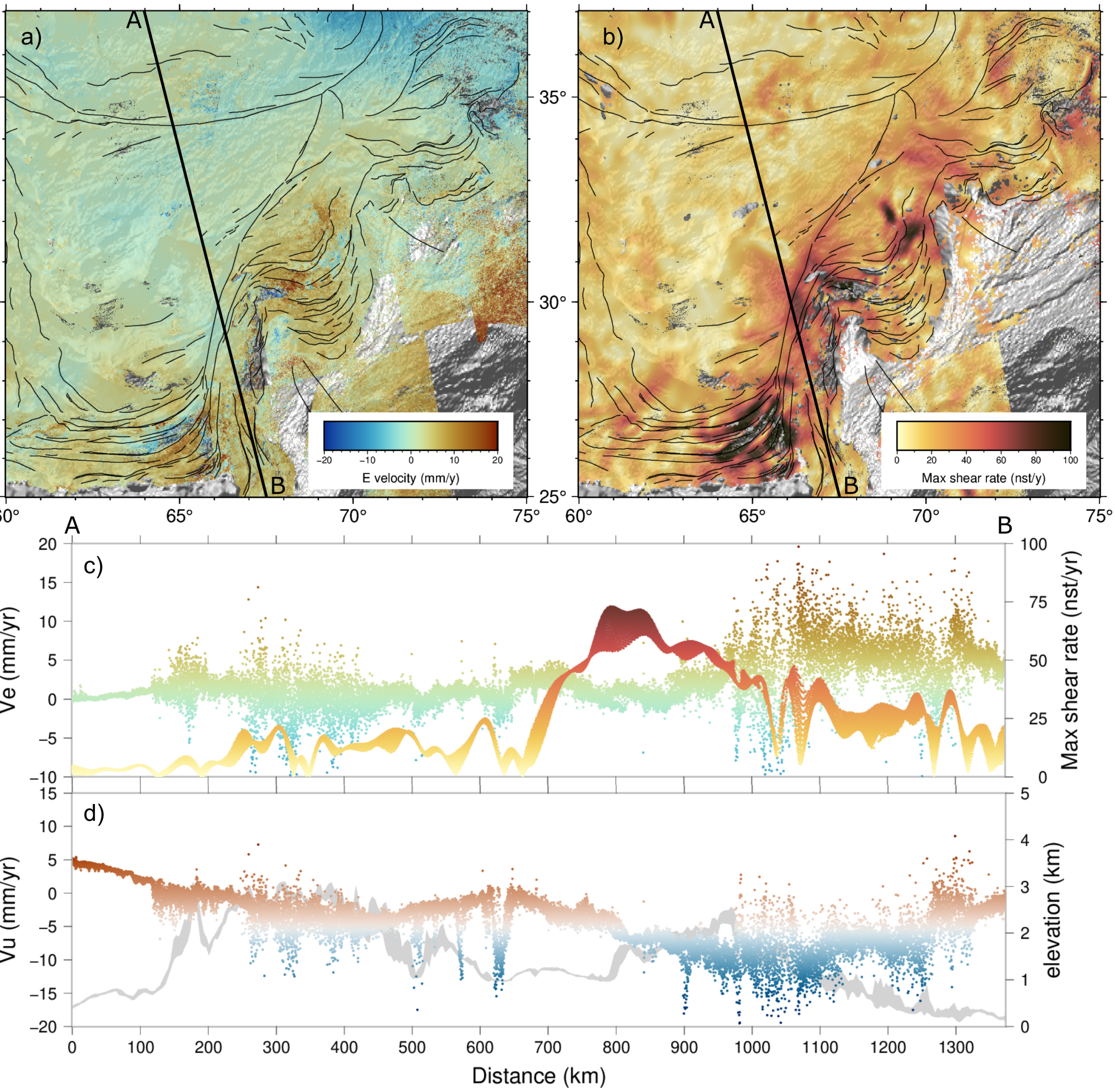

Figure S3: Velocity and strain rate maps and profiles for the region of the Chaman Fault Zone form the Makran, though the Sulaiman Ranges to the Hindu Kush: a) East Velocity; b) Maximum Shear Strain Rate; c) profile through eastwards velocity and maximum shear strain rate fields between points A and B; d) profile from vertical velocity and elevation between A and B.

2.4. Pamirs

The Pamirs lie between the Afghan-Tajik Depression (McNab et al., 2019) in the west and the Tarim basin in the east, and they have formed northwards of the western syntaxis of the Himalayas as a result of the northward motion of India. In addition to a large degree of crustal shortening and associated seismicity, this region is also unusual along the Alpine Himalayan Belt in having intermediate depth earthquakes as well (Schurr et al., 2014). The current shortening across the Pamirs is known to be largely accommodated on the Pamir Thrust system (Ischuk et al., 2013) and north of this major fault zone lies the Tian Shan (Section 2.5).

Major earthquakes have recently struck the region including the 2015 M_w 7.2 earthquake in the Central Pamir, Tajikistan, which was a strike-slip rupture of the Sarez-Karakul fault (Metzger et al., 2017; Zubovich et al., 2022) that was part of a sequence of earthquakes that continued around the Pamirs (Bloch et al., 2023). Previous InSAR studies have measured the deformation field (Metzger et al., 2021), but the coherence in the mountainous and snow-covered region is challenging, resulting in largely the Afghan-Tajik Depression being retrieved in their velocity fields.

In terms of our results, the coherence in both the ascending and descending is fairly good (Figure SB9) except for the highest peaks in the Pamirs and to the south towards the Eastern Syntaxis resulting in some gaps in the velocities (Figure 6). The formal uncertainties calculated from the time series and velocity decomposition indicate that the line-of-sight and 3-component rates are around a 1σ level of 1 mm/yr (Figure 5 & SB18). The eastward velocity with respect to Eurasia shows, in agreement with the sparse GNSS velocities, that most of the immediate region outside of the Pamir is moving only in the north-south direction, with little east-west component (Figure S4a). However, within the Pamirs, there is a significant amount of westward motion from the Pamirs into the Afghan-Tajik Depression where the dilatation (Figure S4d) shows a large amount of contraction ($> 100 \times 10^{-9}$ strain/yr) associated with shortening across the thrust faults in this region (McNab et al., 2019).

There is a sharp gradient in this relative westward motion of the Pamirs (right-lateral shear) at the Pamir Thrust where we see the maximum shear strain rate is co-located with the large gradient in north-south GNSS velocities across the Pamir Thrust system. This strain peaks at 150×10^{-9}strain/yr although our velocity pixel spacing at ~1 km and smoothing of the strains means this peak is somewhat subdued and broadened. An anticipated steeper gradient is suggested by the sharp transition in the east-west velocities seen in the north-south transect in Figure S4h that is almost step-like in profile. However, there is also the potential for unwrapping errors in the interferograms along this steep margin where there is a topographic step of almost 4 km kilometres over a distance of less than 20 km. Following the trace of the east-west velocity step and maximum shear westwards towards Dushanbe, the sharp step in velocity follows the Vakhsh thrust confirming this is a region of active fault creep (Metzger et al., 2021), as is the along-strike continuation to the Ilyak fault immediately north of Dushanbe. Continuing around onto the north-south striking Babatag thrust, there is another sharp gradient in velocity imaged across it with a step of 5 mm/yr observed (Figure S4g).

Our uplift results carry much greater uncertainty in terms of interpreting tectonic contributions (rather than formal uncertainties themselves Figure SB18), but we do see systematic patterns of broad uplift in the south and western Pamir of 5 mm/yr (Figure S4b,g,h), which agree with the few 3-component GNSS rates available (albeit with large uncertainties themselves). Regions of apparent subsidence in the Afghan-Tajik Depression are likely associated with agriculture and therefore are probably artefacts due to phase bias (Ansari et al., 2021). The non-agricultural areas within this region associated with folds and exposed bedrock show negligible uplift rates around 0 mm/yr except west of Dushanbe where we see some potential tectonic uplift associated with the folds. Further westwards even stronger values of apparent uplift > 10 mm/yr also closely follow the desert land cover of Western Turkmenistan and it is not possible to discern if these relate to tectonic or near surface processes, or are in fact long wavelength errors in our velocity fields due to a lack of GNSS constraints for this area (Figure SB18).

Whilst our time series starts after (Table SA3) the 2015 November Sarez earthquake (Metzger et al., 2017; Elliott et al., 2020), the postseismic motion following this event (Jin et al., 2022) is also visible within our results, in particular the NE-SW orientated maximum shear strain located south of the Pamir Thrust System. Whilst much smaller in magnitude, some of the maximum shear along the eastern end of the Pamir thrust system will also be attributable to the postseismic motion from the June 2016 M_w 6.4 Sary-Tash and November 2016 M_w 6.6 Muji earthquakes.

2.5. Tian Shan

The Tian Shan is a relatively young mountain belt reactivated as a far-field response to the ongoing collision between India and Eurasia (Jolivet et al., 2010). It is bounded between the Tarim Basin in the south, the Pamir in the southwest, and the Kazakh Platform in the north, which represents the Stable Eurasia. Absorbing over half of the shortening between India and Eurasia (Zubovich et al., 2010), the Tian Shan is seismically active, with over ten magnitude greater than 7 earthquakes that occurred in the mountain range over the past just over a century (Kulikova, 2016), including the 1889 M8 Chilik Earthquake (Krüger et al., 2017), the 1911 M8 Chon-Kemin Earthquake (Kulikova and Krüger, 2015), and the recent 2024 M_w 7.0 Wushi Earthquake that happened on the southern range front (Zheng et al., 2024).

However, unlike the Pamir which concentrates most of its strain on a single structure, the deformation in Tian Shan is largely distributed, as evident in the gentle gradients and almost linear east velocity profiles both perpendicular and parallel to the range (Figure S5a,e,f). In terms of strain, the negative dilatation (contraction) rates are up to twice in magnitude as the maximum shear, consistent with the compressional regime that dominates Tian Shan's deformation. The spatial patterns of the maximum shear and the dilatation strains are rather similar (Figure S5c,d), and are all slightly more concentrated in the western Tian Shan. However, within this region they are still fairly evenly distributed and rarely exceed 50 nstr/yr. This is consistent with geological observations that show the shortening in the western Tian Shan being partitioned between slip rates on the range-frontal folds-and-thrust belts, basin-bounding faults and basin-interior folds (Thompson et al., 2002). Hence, the deformation in Tian Shan is more similar to that described by a continuum model (England and Molnar, 2015). Elsewhere in the Tian Shan, the strain rates are mostly below 30 nstr/yr, which is around the noise level of our measurements (Section 2.3).

Apart from the range-parallel thrusts and folds, the Tian Shan is thought to accommodate shortening via the northwest-trending right-lateral strike-slip faults that extend into the crystalline Kazakh Platform. The Talas-Fergana fault and the Dzhungarian fault, in particular, have pronounced geomorphological features that suggest they have ruptured in very large earthquakes (Burtman et al., 1996; Campbell et al., 2015). However, in our datasets, we do not observe significant east velocity steps across these two structures, or any of the northwest-trending strike-slip faults. The only signal that might suggest the role played by the strike-slip faulting is the slightly higher shear strain along the segment of the Talas Fergana fault dividing the Fergana Basin and the western Tian Shan (Figure S5c). However, the magnitude of shear rate on this fault is not higher than that in the interior western Tian Shan, suggesting that its role is minor. This is consistent with published geodetic slip rates on the slow end (Zubovich et al., 2010; Wu et al., 2023), in contrast to the often fast slip rates derived from geological methods (Rizza et al., 2019; Rust et al., 2018).

The vertical velocity field shows gentle uplift across the Tian Shan, punctuated by subsidence signals from agriculture, mining, oil and gas exploitation, and groundwater extraction with likely impact from phase bias (Ansari et al., 2021). The rapid subsidence might also have introduced artifacts into the vertical velocities in regions with sparse vertical GNSS control, as VELMAP inherently assumes zero average vertical velocities at mesh nodes during the inversion for referencing the LOS velocities (Wang and Wright, 2012). This might have implications for interpreting the ~5 mm/yr uplift along the Borohoro Mountain, and into the Kazakh Platform, for example. It is likely that the high uplift rates between 1200 and 1600 km along the profile B-B' are overestimated.

The circle of gentle subsidence around the Lake Issyk-Kul might be real. This subsidence could be due to permafrost thaw, or the delamination of the lithosphere underneath (Pan et al., 2019), or a combination of both. Just to the east of the Issyk-Kul lake is the Karkara Range front Fault which is thought to be creeping at rates up to 3 mm/yr (Mackenzie et al., 2018). However, no obvious contraction nor uplift was resolved from our data sets, possibly because the motion is largely north-south oriented, or because it is creeping at a rate lower than the noise level of our velocities.

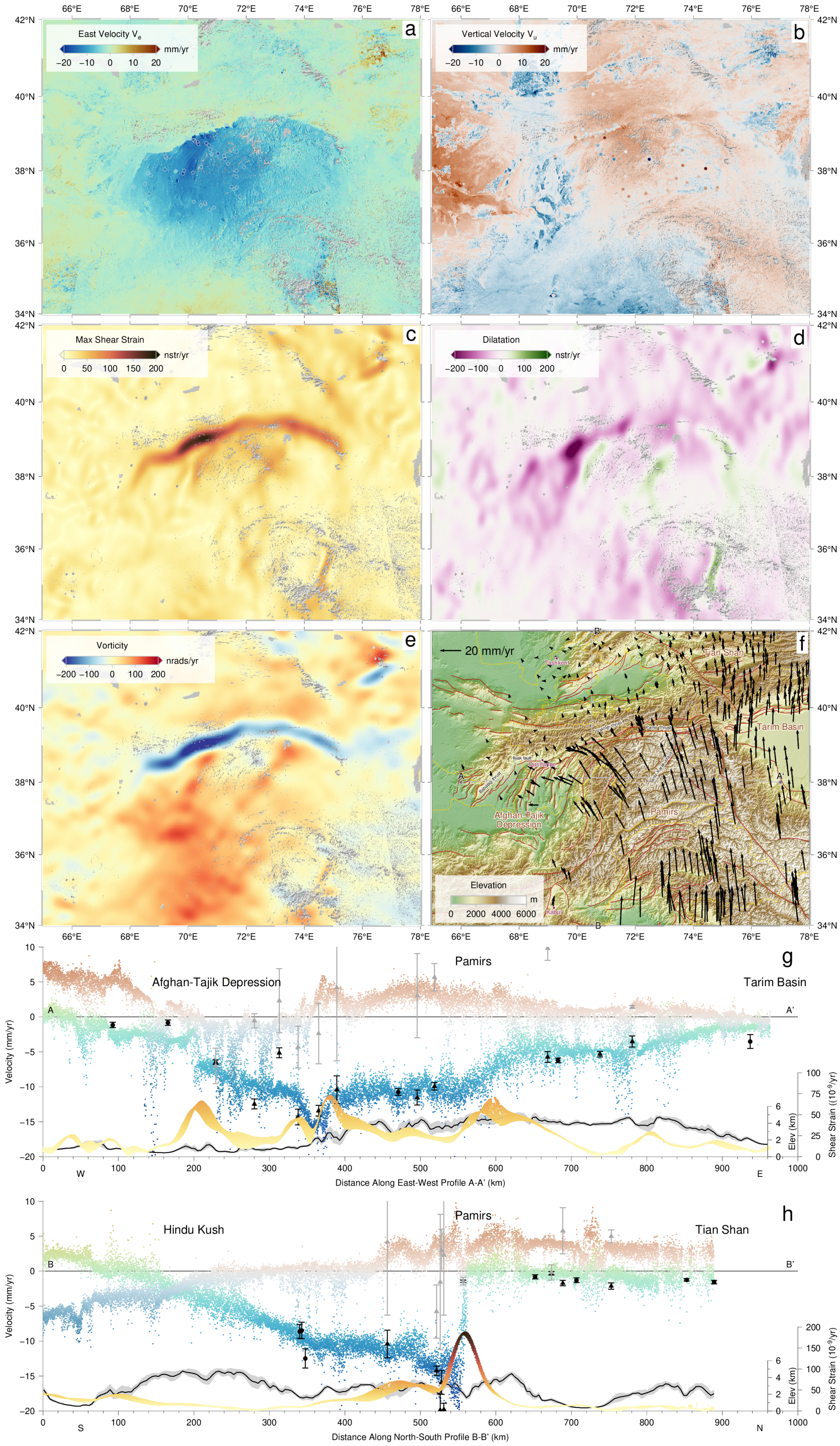

Figure S4: Velocity and strain rate maps and profiles for the region of the Pamirs. (a) Component of Eastward Velocity with GNSS eastwrad velocities as coloured circles, (b) Component of Vertical Velocity with 3D GNSS vertical component as coloured circles, (c) Maximum Shear Strain Rate, (d) Horizontal Dilatation Rate (positive is extension), (e) Vorticity (positive is anticlockwise rotation), (f) Elevation map of the Pamirs with GNSS velocities overlaid (black arrows), active faults from (Styron and Pagani, 2020) and location of profiles in grey, (g) East-west profile A–A' through velocity and strain rate fields (with GNSS and topography marked). (h) North-south profile B–B'.

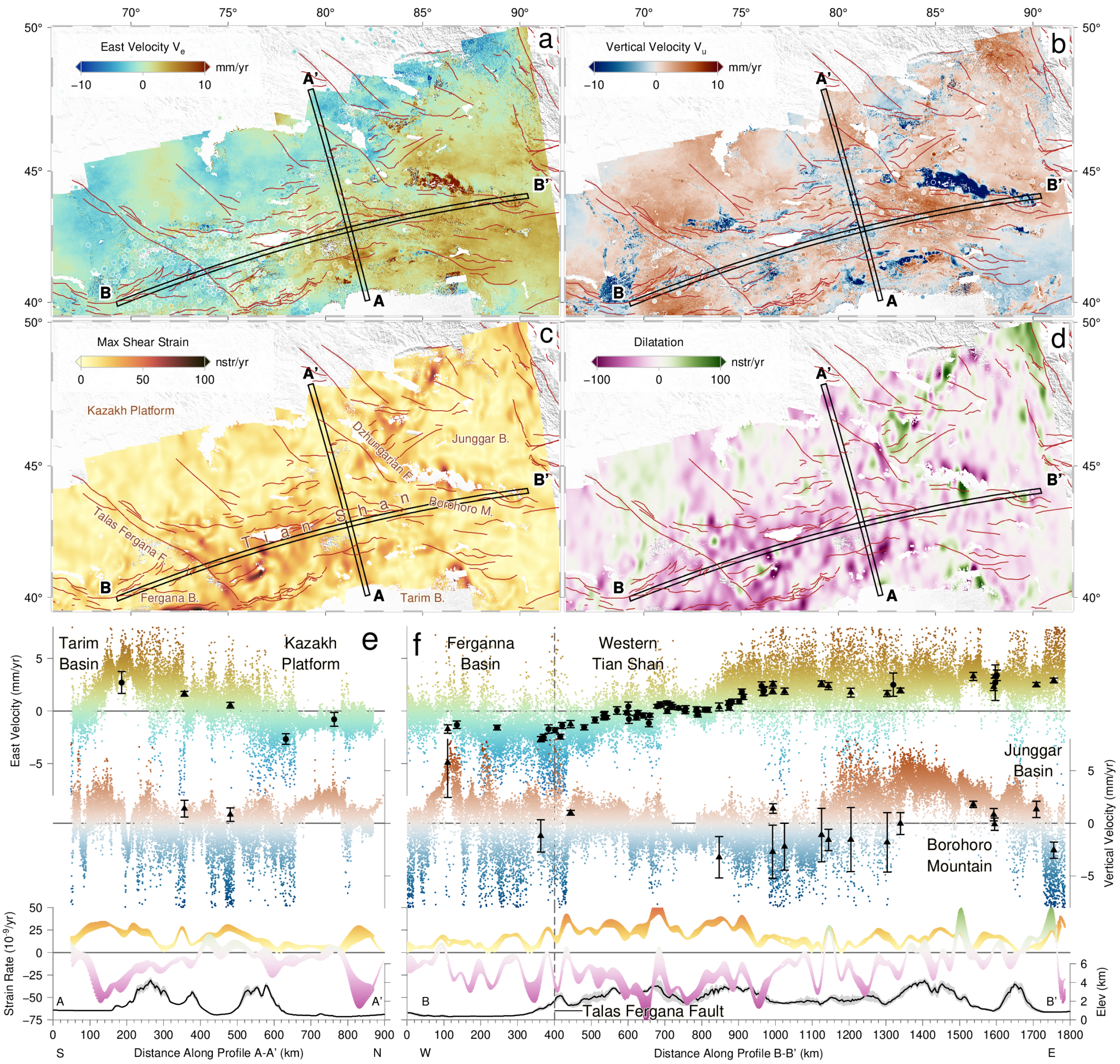

Figure S5: Velocity and strain rate maps and profiles for the region of the Tian Shan. (a) Component of Eastward Velocity with GNSS eastward velocities as coloured circles, (b) Component of Vertical Velocity with 3D GNSS vertical component as coloured circles, (c) Maximum Shear Strain Rate, (d) Horizontal Dilatation Rate (positive is extension), (e) North-south profile A–A' through velocity and strain rate fields (with GNSS and topography marked). (f) East-west Profile B-B'. B.=Basin. F.=Fault. M.=Mountain. Active faults from Styron and Pagani (2020).

2.6. *Western Tibet*

Taking 98°E as a boundary, we can subdivide the Tibetan Plateau into Western Tibet and Eastern Tibet. Northwest of Western Tibet lies the Tarim basin, a relatively rigid block with low topography, little internal crust deformation and minimal seismic activity (Avouac and Tapponnier, 1993). To accommodate ~50 mm/yr of collision between India and Eurasia (Molnar and Tapponnier, 1975), numerous active faults have formed in Western Tibet, with the Altyn Tagh Fault being the most significant. This fault is one of the longest active strike-slip faults in the world, with a length of ~1500 km (Tapponnier and Molnar, 1977), and acts as the boundary between Tibet and the Tarim basin (Molnar and Dayem, 2010). Besides the Altyn Tagh fault, other major active faults in Western Tibet include the Karakoram fault (e.g. Avouac and Tapponnier (1993); Wright et al. (2004)), the Karakax fault ((e.g., Peltzer et al., 2020)), the Kunlun fault ((e.g., Kirby et al., 2007; Harkins et al., 2010; Diao et al., 2019)) and the Manyi fault ((e.g., Bell et al., 2011)). Several active faults are also found both inside and around the Qaidam basin ((e.g., Daout et al., 2019)), as well as in the Qilian Shan - Nan Shan thrust belt ((e.g., Hao et al., 2015; Xu et al., 2021; Wang et al., 2024a)).

Strong earthquakes within the Western Tibet during the last 100 years include the 1932 M_w 7.6 Changma Earthquake on the Changma Fault (Du et al., 2019), the 2008 M_w 7.2 Yutian Earthquake and the 2014 M_w 6.9 Yutian Earthquake on the Longmu-Gozha Co fault system (Li et al., 2015, 2016, 2020), the 1997 M_w 7.5 Manyi Earthquake on the Manyi fault (Funning et al., 2007; Wang et al., 2007) and the 2001 M_w 7.8 Kokoxili Earthquake on the Kunlun fault (Lasserre et al., 2005). It should be noted that there has been an absence of major earthquakes for over 300 years on the central Altyn Tagh Fault (between ~ 84°E and ~ 94°E (Yuan et al., 2020).

Previous GNSS and InSAR studies over the Western Tibet are mainly focused on fault kinematic parameter estimation, especially the slip rate of the Altyn Tagh fault (e.g., Bendick et al., 2000; Mériaux et al., 2004; Wallace et al., 2004; Zhang et al., 2007; Elliott et al., 2008; Li et al., 2018; Shen et al., 2019), as this can help us better understand how the Tibet Plateau accommodates the strain generated by the collision between India and Eurasia (e.g., Tapponnier et al., 2001; England and Molnar, 2005). The GNSS velocity field has short comings in determining the locking depth of faults and sources of strain as it is normally too sparse in this region; however, the InSAR velocities, especially the large-scale InSAR velocities that merge several tracks ((e.g., Shen et al., 2024; Wang et al., 2024a)), can provide a good opportunity to study strain localization and spatial variation of fault kinematic parameters in the Western Tibet.

Due to the good coherence in this region (Figure SB9), the velocity field and strain rate results (Figure 6, Figure 7) show a good coverage over Western Tibet, and as we obtained our LOS velocities based on ~8-year time series, the uncertainties of the long-term rates are at a relatively low level, with most of them smaller than 1 mm/yr based on the LiCSBAS time-series analysis (Figure 5). The north velocities of Western Tibet show relatively low uncertainty in most of the region, and only show large uncertainties of over 2 mm/yr around the western Kunlun fault and the north side of the Altyn Tagh fault due to the missing of GPS velocities (Figure 5). The results we obtained revealed some key surface deformation characteristics of this region. By zooming in on this area (Figure S6), we could see that our decomposed east velocities fit the GNSS east velocities (the coloured circles in Figure S6a) very well. We found the east-west velocity field pattern is mainly controlled by the left-lateral strike-slip motion along the active strike-slip faults (e.g., the Altyn Tagh Fault, the Kunlun Fault, and the Manyi Fault) in this region (Figure S6a) and that high strain rates mainly focus on these major active faults (Figure S6b).

We see maximum shear strain rates of over 50 nst/yr along the Altyn Tagh fault between 84°E and 91°E and relatively low maximum shear strain rate between 92°E and 95°E (Figure S6b), which is consistent with previous research and can be explained by a strike-slip rate of near 8 mm/yr on the deep shearing part of the fault between 84°E and 91°E and a lower strike-slip rate of near 5 mm/yr between 92°E and 95°E (Wang et al., 2024a). The high maximum shear strain rates on the Altyn Tagh Fault do not continue further east than 95°E, supporting the view that the strain accumulation on the Altyn Tagh Fault is terminated around this area (Wang et al., 2024a). The Altyn Tagh fault shows relatively higher shear rates at 90.5°E (around the Pingding Shan bend), which may be related to shallow creeping indicated by previous studies (e.g. Wang et al. (2024a)).

We also found very high maximum shear strain rates of over 100 nst/yr along the western Altyn Tagh Fault (or the Longmu-Gozha Co fault system), the Kunlun Fault and the Manyi Fault (Figure S6b,e,f,h,i). Previous studies (Li et al., 2015, 2016, 2020; Funning et al., 2007; Wang et al., 2007; Lasserre et al., 2005) show that all these areas experienced strong earthquakes with a magnitude of over 6.5 during the last 30 years, which include the 2008 M_w 7.2 Yutian Earthquake, the 2014 M_w 6.9 Yutian Earthquake, the 1997 M_w 7.5 Manyi Earthquake and the 2001 M_w 7.8 Kokoxili Earthquake. Therefore, the high strain rates we observe around these places are at least in part due to postseismic deformation related to these strong earthquakes (e.g., Huang et al., 2023; Ryder et al., 2007; Yamasaki and Houseman, 2012; Ryder et al., 2011; Wen et al., 2012).

The dilatation rates (Figure S6c) indicate that there is distributed compression around the Altyn Tagh fault. Other regions such as the Himalayas, the Qaidam basin and the Qilian Shan - Nan Shan Thrust belt also exhibit distributed compression. This pattern is consistent with previous studies (e.g., Avouac and Tapponnier, 1993; Yin et al., 2008; Wang et al., 2024a), suggesting that these areas are the main thrust zones in western Tibet. Distributed extension (positive dilatation rates) is observed in Central Tibet (Figure S6c), which is consistent with the widely distributed N-S trending normal faults present in this region (Tapponnier et al., 1981; Wang et al., 2010). The Vorticity map (Figure S6d) can be used to show the rotation direction and rates of the crust in Western Tibet, but it will also capture the strike-slip motion along the faults. The Altyn Tagh fault and the Kunlun fault show positive vorticity rates due to the left-lateral motion of crust along them. Central Tibet exhibits a noisy pattern in the vorticity map, which may be due to the widely distributed faults and sub-blocks within the region. We can observe negative vorticity rates around the Tarim basin, supporting a clockwise rotation of the rigid Tarim that indicated by previous studies (e.g., Wang and Shen, 2020).

Vertical velocities are also given in our studies (Figure 7). Unlike the East velocities which are dominantly controlled by tectonic motion, the vertical velocities are more sensitive to non-tectonic activities such as permafrost, hydrology and anthropogenic activities (e.g., Gruber, 2012; Ou et al., 2022; Wang et al., 2024a). Moreover, the remaining phase bias (e.g., Ansari et al., 2021; Maghsoudi et al., 2022) also maps primarily into the vertical velocity field. Although the vertical velocity field is influenced by various non-tectonic sources, the long-wavelength signals in it can still reflects some tectonic activities within Western Tibet. Strong shortening and thrust motion due to India-Eurasia collision generates rapid uplift rates that we observe around Himalayas. Distributed uplifts are also observed within the Qaidam basin and the Qilian Shan - Nan Shan thrust belts, which may be controlled by the thrust motion along the active faults within them (Wang et al., 2024a).

2.7. *Eastern Tibet*

Eastern Tibet is a key region within the AHB, situated between the high-relief plateau interior to the west and the stable Sichuan Basin to the east, with the eastern Himalayan syntaxis and the Indo-Burma subduction zone to the south. This region is a transition zone where the internal deformation of the Tibetan Plateau is accommodated by large-scale strike-slip faulting along major faults (i.e., Haiyuan, Kunlun, and Xianshuihe Faults) and distributed off-fault deformation (Fang et al., 2024b). The driving forces behind the deformation include the ongoing convergence of India and Eurasia, active faulting resistant to deformation, and gravitational potential energy (GPE) from the high plateau topography (Houseman and England, 1986; Flesch et al., 2001; Fang et al., 2024a).

Eastern Tibet has experienced several large earthquakes, highlighting its active tectonics. Notable events include the 2008 M_w 7.9 Wenchuan (Qi et al., 2011), 2013 M_w 6.6 Lushan (Xu et al., 2013), 2017 M_w 6.5 Jiuzhaigou (Sun et al., 2018), 2021 M_w 7.4 Maduo (Fang et al., 2022), 2021 M_w 6.1 Yangbi (Zhou et al., 2022), 2022 M_w 6.6 Menyuan (Yang et al., 2023), 2022 M_w 6.7 Luding (Guo et al., 2023), and 2025 M_w 7.9 Myanmar earthquakes.

Previous studies have quantified surface deformation and strain partitioning in eastern Tibet using GNSS and InSAR. Shen et al. (2005) derived a detailed horizontal velocity field for southeastern Tibet from GNSS. Wang and Shen (2020) updated the deformation field with higher-density GNSS networks, refining strain rate estimates. Wang et al. (2021) analyzed fault slip rates and regional kinematics from GNSS observations. More recently, InSAR time series analysis has been used to resolve interseismic strain accumulation (Qiao and Zhou, 2021; Huang and Zhou, 2022; Zhao et al., 2024; Fang et al., 2024b).

Our analysis benefits from extensive Sentinel-1 InSAR coverage in eastern Tibet. The coherence of the interferograms varies spatially, influenced by surface properties and temporal decorrelation effects. Coherence is lower in the southern part of the study area, where dense vegetation causes significant temporal decorrelation (Figure SB9), leading to relatively higher uncertainties (3-5 mm/yr) in the velocities (Figure SB18). Overall, the InSAR-derived velocities show a good fit to GNSS data, with a root mean square (RMS) difference of approximately 2 mm/yr.

The E-W velocity field reveals clear contrasts across major faults, especially creeping segments of the Haiyuan and Xianshuihe Faults, where strain localization is observed, with a peak of approximately 100 nanostrain/yr (Figure S7). This value is likely underestimated due to the Gaussian filtering applied to the east velocities, which smooths out some of the finer-scale deformation signals across the faults. Apart from these localized features, both diffuse extension across the high plateau and contraction in northeastern Tibet, not directly associated with mapped known faults, can be explained by distributed moment sources (Fang et al., 2024b). The velocity field also shows anticlockwise rotation on the major faults, while the high plateau exhibits clockwise rotation around the eastern Himalayan syntaxis (EHS) (Figure S7d), consistent with the horizontal flow revealed by GNSS observations (Figure S7a). It is important to note that the ongoing postseismic deformation from the 2008 M_w 7.9 Wenchuan earthquake has caused apparent strain concentration, contraction (Figure S7b-c), and uplifting (Figure 7c) along the Longmenshan Fault.

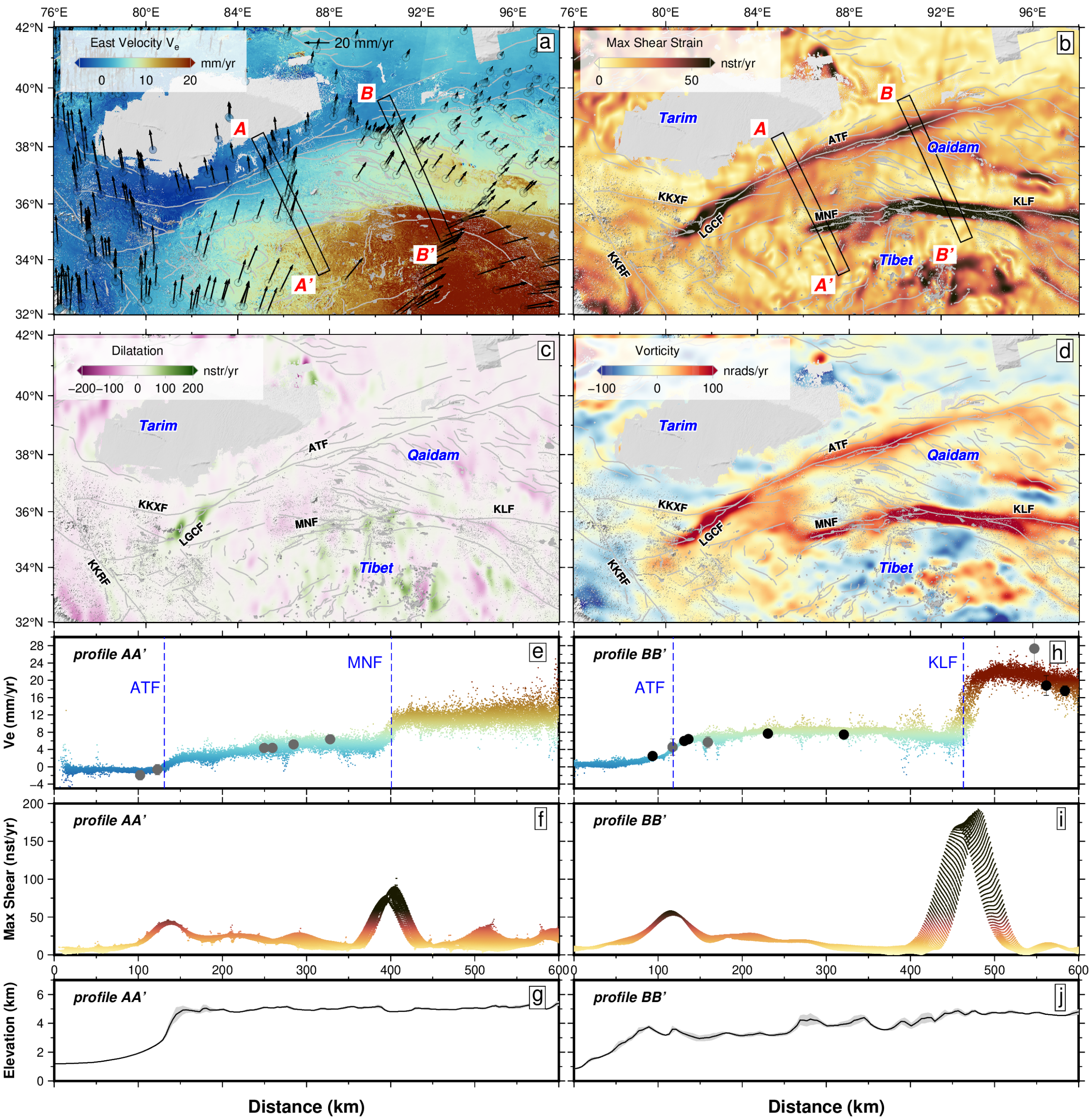

Figure S6: Velocity and strain rate maps and profiles for the region of Northern and West Tibet. (a) East Velocity. The GNSS east velocities are shown by coloured circles, and the GNSS horizontal velocities are shown by black arrows. (b) Maximum Shear strain rates. (c) Dilatation, with positive values indicating extension. (d) Vorticity, with positive indicating anti-clockwise rotation. (e–g) The eastward velocities, maximum shear strain rates and elevation for profile A–A' in (a), black and gray circles in (e) represent the east components of 3D and 2D GNSS velocities, respectively. (h-j) The eastward velocities, maximum shear strain rates and elevation for profile B–B' in (a), black and gray circles in (h) represent the east components of 3D and 2D GNSS velocities, respectively. ATF: Altyn Tagh Fault, MNF: Manyi Fault, KKXF: Karakax fault, KKRF: Karakoram fault, LGCF: Longmu-Gozha Co fault.

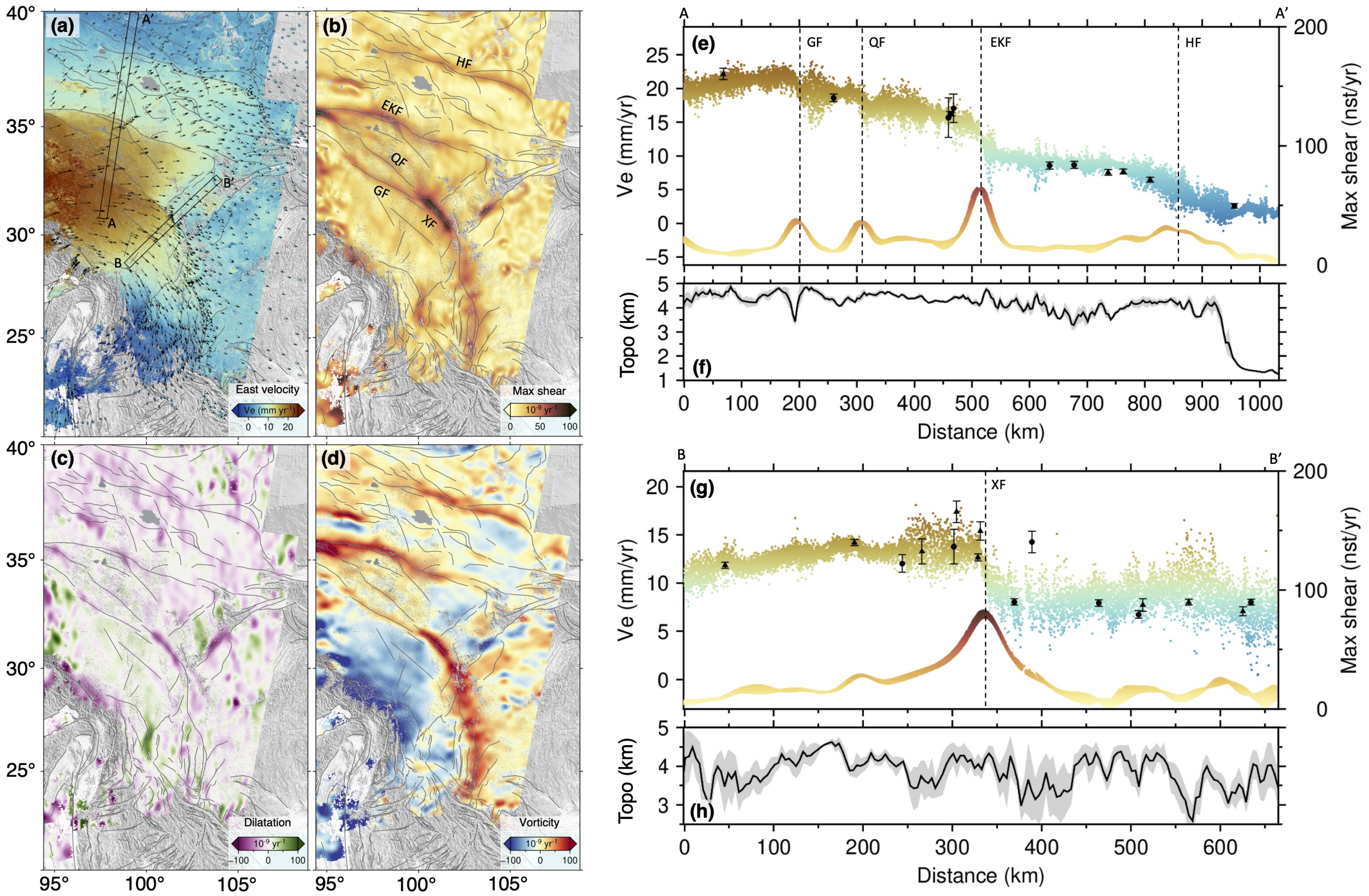

Figure S7: Velocity and strain rate maps and profiles for the region of Eastern Tibet. (a) Eastward velocity. GNSS eastward velocities are represented by coloured circles (2D sites) and triangles (3D sites). Arrows indicate horizontal GNSS velocities. The rectangle marks the location of Profiles A–A' and B-B' (20 km width), shown in (e-h). (b) Maximum shear strain rate. (c) Dilatation, with positive values indicating extension. (d) Vorticity, with positive indicating anti-clockwise rotation. (e–f) Profile A-A' through eastward velocity and maximum shear strain rate, with GNSS eastward velocities (40 km bin), faults, and topography marked on. (g-h) Profile B-B'. EKF = East Kunlun Fault, GF = Ganzi Fault, HF = Haiyuan Fault, QF = Qingshuihe Fault, XSHF = Xianshuihe Fault.

Appendix A. Supplementary Tables

Reference	Number of Sites	Sites Used	Region	
Abbasi et al. (2023)	20	20	57.3–62.3°E	25.2–28.3°N
Ader et al. (2012)	34	34	80.5–87.7°E	26.4–32.0°N
Aktug et al. (2016)	142	30	33.4–56.2°E	26.2–39.1°N
Alothman et al. (2016) (DORIS)	14	7	38.8–49.0°E	23.5–29.1°N
Altamimi et al. (2016) (SLR)	20	2	86.8–86.8°E	28.0–28.0°N
Altamimi et al. (2016) (VLBI)	66	8	20.7–82.2°E	35.4–51.3°N
Ashurkov et al. (2016)	4	4	103.1–107.6°E	51.7–51.8°N
Atanasova-Zlatareva (2014)	30	9	22.7–28.4°E	41.6–44.1°N
Barman et al. (2017)	39	28	88.6–93.9°E	23.7–27.6°N
Bisht et al. (2021)	6	6	77.2–93.8°E	27.4–31.8°N
Bitharis et al. (2024)	207	184	7.5–41.6°E	30.6–49.1°N
Blewitt et al. (2018)	3744	555	0.3–119.9°E	20.3–51.9°N
Bock et al. (2021)	238	217	0.3–118.4°E	22.2–52.0°N
Briole et al. (2021)	327	318	19.9–33.9°E	35.0–43.1°N
Brockmann et al. (2019)	2514	1973	0.0–39.8°E	30.6–52.0°N
Castro-Perdomo et al. (2022)	102	96	7.5–91.1°E	22.2–50.4°N
Cheloni et al. (2014)	147	3	10.3–12.5°E	45.5–46.0°N
Choudhury et al. (2018)	1	1	72.7–72.7°E	23.2–23.2°N
Chousianitis et al. (2015)	99	18	19.8–24.0°E	36.8–39.8°N
d'Agostino et al. (2020)	336	326	1.0–31.3°E	35.1–52.0°N
Dai et al. (2023)	28	24	99.4–100.4°E	25.6–26.5°N
Dal Zilio et al. (2020)	92	92	85.0–86.0°E	27.3–28.0°N
DeMets et al. (2020)	56	48	0.3–87.3°E	20.3–51.0°N
Devachandra et al. (2014)	17	10	95.8–97.0°E	27.8–28.8°N
Diao et al. (2019)	27	24	97.5–101.1°E	34.0–36.7°N
Dimitrov and Nakov (2022)	28	26	22.7–24.1°E	41.6–43.1°N
Duong et al. (2013)	22	22	103.0–105.3°E	20.6–22.3°N
England et al. (2016)	346	115	19.5–39.7°E	35.4–41.8°N
Ergeshov et al. (2020)	6	2	60.7–67.0°E	37.5–41.5°N
Ergintav et al. (2014)	112	5	26.6–29.2°E	39.9–41.3°N
Ergintav et al. (2023)	796	720	19.5–44.0°E	33.7–43.0°N
FANG et al. (2022)	94	94	74.1–94.4°E	23.1–34.6°N
Fazilova et al. (2018)	5	2	66.9–70.6°E	38.7–42.0°N
Fazilova et al. (2025b)	16	13	68.7–70.6°E	40.3–42.0°N
Fazilova et al. (2025a)	9	9	65.9–66.0°E	38.5–38.6°N
Floyd et al. (2023)	360	332	19.5–30.9°E	34.8–42.0°N
Frohling and Szeliga (2016)	19	11	62.3–67.1°E	24.9–29.1°N
Fu and Freymueller (2012)	25	24	80.5–91.1°E	26.4–30.0°N
Fu et al. (2020)	15	12	102.7–103.3°E	26.1–27.1°N
Gahalaut et al. (2019)	22	22	69.0–72.8°E	21.0–24.7°N
Gan et al. (2022)	501	456	95.5–107.1°E	20.7–34.0°N
Gautam et al. (2017)	6	5	77.1–80.2°E	28.5–30.7°N
Gorshkov et al. (2023) (LSS)	200	179	12.9–60.0°E	41.6–51.9°N
Gorshkov et al. (2023) (MED)	200	173	12.9–60.0°E	41.6–51.9°N
Gülal et al. (2013)	40	1	29.8–29.8°E	37.5–37.5°N
Gupta et al. (2015)	37	26	90.3–96.5°E	26.4–28.8°N
Hamiel and Piatibratova (2021)	290	284	15.0–51.3°E	24.9–44.4°N
Hao et al. (2016)	38	38	97.2–104.2°E	21.9–31.6°N
Hao et al. (2019)	928	903	74.3–120.0°E	20.0–49.6°N
Hao et al. (2021)	990	970	101.0–117.0°E	31.0–43.9°N
Hao et al. (2024)	177	176	99.0–112.2°E	33.6–42.3°N
Huang et al. (2023)	26	26	100.5–102.7°E	23.0–25.3°N
Jade et al. (2011)	18	10	77.2–79.6°E	28.5–34.6°N
Jade et al. (2014)	67	52	76.3–92.4°E	27.3–34.6°N
Jade et al. (2017)	68	37	50.6–102.8°E	20.3–43.2°N
Jade et al. (2020)	39	35	66.9–114.4°E	20.3–43.2°N
Jolivet et al. (2023)	39	32	30.5–35.5°E	39.4–42.0°N
Jouanne et al. (2017)	78	76	80.1–87.7°E	26.5–30.0°N
Jouanne et al. (2020)	65	64	71.6–77.5°E	31.8–36.8°N
Kadirov et al. (2014)	86	74	43.8–49.4°E	37.8–43.0°N
Kannaujiya et al. (2022)	15	15	77.6–79.9°E	29.0–30.7°N
Ghasemi Khalkhali et al. (2021)	25	25	44.2–47.7°E	36.9–39.4°N
Ghasemi Khalkhali et al. (2023a)	31	31	54.4–61.7°E	30.8–37.8°N
Ghasemi Khalkhali et al. (2023b)	10	10	55.1–62.3°E	25.3–31.3°N
Khorrami et al. (2019)	396	338	39.8–61.7°E	23.6–44.9°N
Kreemer et al. (2014)	7374	2397	9.1–119.6°E	20.3–51.8°N
Kufner et al. (2021)	20	20	68.6–71.8°E	35.3–38.0°N
Kumar et al. (2023)	9	9	75.4–77.4°E	30.7–32.5°N
Kurt et al. (2023)	836	829	25.9–44.6°E	36.0–42.0°N
Li et al. (2017)	212	198	100.1–107.4°E	33.9–39.4°N
Li et al. (2018)	994	26	85.5–90.8°E	35.4–39.0°N
Li et al. (2021)	200	194	100.1–107.4°E	33.9–39.4°N
Li et al. (2022)	73	69	75.7–79.4°E	39.5–40.9°N
Continued in the next Table...				

Table SA1: GNSS studies used in this compilation with the original number of site velocities in this study and the number remaining after analysis. The latitudinal and longitudinal ranges are the regions covered by the velocities included here. DORIS - Doppler Orbitography and Radiopositioning by Integrated Satellite. SLR - Satellite Laser Ranging. VLBI - Very Long Baseline Interferometry. Continued in next Table SA2.

Reference	Number of Sites	Sites Used	Region	
Continued from previous Table...				
Lindsey et al. (2023)	68	64	89.5–97.5°E	20.0–27.3°N
Mallick et al. (2019)	20	20	88.7–96.1°E	22.4–27.3°N
Mantovani et al. (2015)	487	436	5.4–26.1°E	36.7–48.4°N
Marechal et al. (2016)	35	24	89.4–91.5°E	26.9–28.1°N
Metzger et al. (2020)	74	70	68.4–74.8°E	36.7–40.8°N
Meyer et al. (2024)	28	26	71.6–73.9°E	31.8–34.0°N
Milyukov et al. (2015)	7	3	42.5–44.7°E	43.0–43.7°N
Milyukov et al. (2022)	13	12	38.1–42.0°E	43.6–47.9°N
Mironov et al. (2021)	10	10	41.6–44.7°E	42.4–43.8°N
Mohanty et al. (2024)	32	15	76.2–79.3°E	32.8–35.9°N
Mukul et al. (2018)	24	16	87.3–89.2°E	23.5–28.0°N
Mullick and Mukhopadhyay (2020)	51	50	86.4–92.9°E	22.1–26.5°N
Nagale et al. (2022)	18	15	81.0–91.1°E	26.5–29.7°N
Nocquet et al. (2016)	76	76	3.9–9.9°E	43.0–47.0°N
Osman Pour *in prep.*	57	40	50.9–52.6°E	35.1–36.8°N
Özarpacı et al. (2021)	49	44	29.4–30.9°E	40.1–41.2°N
Özdemir and Karslıoğlu (2019)	206	192	7.5–74.7°E	30.6–50.4°N
Özkan et al. (2023)	58	48	26.1–41.6°E	35.9–44.5°N
Palano et al. (2018)	805	392	34.2–62.2°E	21.4–43.8°N
Pan et al. (2019)	278	274	66.9–94.9°E	35.1–48.7°N
Pan et al. (2020)	272	261	93.0–107.0°E	34.0–40.7°N
Pan et al. (2021)	231	231	74.7–119.8°E	20.9–47.9°N
Pan et al. (2023)	52	52	66.9–89.5°E	36.4–49.8°N
Pappachen et al. (2021)	30	27	44.4–115.9°E	24.9–43.8°N
Perry et al. (2019)	77	8	67.1–79.0°E	24.9–39.6°N
Piña-Valdés et al. (2022)	2853	2788	0.0–44.5°E	30.6–52.0°N
Pintori et al. (2022)	543	543	0.2–20.3°E	42.4–49.8°N
Ponraj et al. (2019)	17	12	79.1–80.6°E	28.6–30.0°N
Qiao et al. (2017)	150	82	74.1–80.5°E	39.4–42.0°N
Rakhimberdieva et al. (2023)	6	1	60.7–60.7°E	41.5–41.5°N
Rudenko et al. (2013)	39	39	0.7–115.9°E	25.0–51.4°N
Rui and Stamps (2019)	1659	1553	50.6–120.0°E	20.3–51.7°N
Santamaría-Gómez et al. (2012)	37	37	0.7–115.9°E	22.2–51.4°N
Serpelloni et al. (2013)	666	664	0.1–36.2°E	29.5–52.0°N
Serpelloni et al. (2016)	475	450	0.3–22.6°E	42.2–52.0°N
Serpelloni et al. (2022)	2336	2062	0.1–116.2°E	21.4–52.0°N
Sharma et al. (2020)	41	39	75.7–80.2°E	26.5–32.4°N
Shirazinejad (*in prep.*)	25	24	44.2–47.4°E	36.9–39.4°N
Shrungeshwara et al. (2023)	13	13	74.3–76.2°E	33.6–34.6°N
Shukurov	11	1	71.8–71.8°E	40.4–40.4°N
Singh et al. (2022)	18	18	73.9–76.2°E	32.7–34.6°N
Socquet and Janex (2019)	816	736	0.1–41.6°E	30.6–52.0°N
Sokhadze et al. (2018)	69	64	15.0–74.7°E	30.6–49.6°N
Steckler et al. (2016)	49	29	90.2–97.4°E	22.4–25.2°N
Su et al. (2018)	58	53	100.1–109.3°E	33.3–38.6°N
Suribabu et al. (2022)	8	7	69.2–70.1°E	23.1–23.2°N
Vardić et al. (2022)	94	83	0.2–114.5°E	22.2–52.0°N
Verma et al. (2024)	22	22	78.9–79.8°E	29.1–29.7°N
Vila (2020)	65	63	80.5–94.8°E	22.4–30.0°N
Vilayev et al. (2017)	10	9	76.0–78.4°E	43.0–43.9°N
Viltres et al. (2022)	149	145	34.7–58.2°E	20.0–36.2°N
Wang et al. (2022b)	618	452	97.5–106.4°E	21.6–34.0°N
Wang et al. (2017a)	29	29	111.2–118.3°E	34.5–42.2°N
Wang et al. (2017b)	1838	1742	65.8–119.8°E	20.3–49.9°N
Wang and Shen (2020)	2370	2288	6.9–120.0°E	20.0–51.8°N
Wang et al. (2020)	524	269	74.3–98.9°E	34.4–44.0°N
Wang et al. (2022a)	765	765	107.1–120.0°E	28.0–42.0°N
Wang et al. (2024c)	546	454	67.0–116.6°E	24.7–46.2°N
Wang et al. (2018)	2	2	51.3–51.3°E	35.7–35.7°N
Wang et al. (2024b)	95	95	85.5–99.1°E	35.4–41.9°N
Wu et al. (2024)	9166	9166	74.7–120.0°E	20.0–49.7°N
Xiong et al. (2021)	1816	1780	74.3–120.0°E	20.0–50.4°N
Yadav et al. (2019)	34	34	78.0–82.2°E	28.7–31.0°N
Yadav et al. (2021)	6	5	78.0–91.1°E	26.9–30.5°N
Yavasoglu et al. (2021)	51	27	26.4–30.0°E	39.7–41.7°N
Younes (2023)	48	36	25.2–34.9°E	22.5–31.5°N
Zhang et al. (2021)	151	145	75.4–98.7°E	20.8–34.8°N
Zhao et al. (2017)	723	672	103.1–117.0°E	32.0–43.9°N
Zhao et al. (2023)	852	852	74.3–110.3°E	21.0–48.7°N
Zheng et al. (2017)	1540	1430	69.2–120.0°E	20.0–49.6°N
Zhou et al. (2016)	149	94	56.1–115.9°E	22.2–43.8°N
Zhou et al. (2023)	25	24	102.6–103.5°E	25.5–27.1°N
Zubovich et al. (2016)	4	4	72.2–72.3°E	39.4–39.6°N

Table SA2: Continued from Table SA1. GNSS studies used in this compilation with the original number of site velocities in this study and the number remaining after analysis. The latitudinal and longitudinal ranges are the region covered by the velocities included here.

Date	Magnitude M_w	Epicentre °E, °N	Country	Timeseries Mitigation
06/02/2023	7.8	37.014, 37.226	Kahramanmaraş, Türkiye/Syria	Ended time-series early
12/11/2017	7.3	45.959, 34.911	Ezgeleh-Sarpolzahab, Iran	Started time-series late
21/05/2021	7.3	98.251, 34.598	Maduo, China	Ended time-series early
30/10/2020	7.0	26.784, 37.897	Aegean Sea, Greece-Türkiye	Solve for coseismic offset
22/01/2024	7.0	78.654, 41.256	Kyrgyzstan-China	Ended time-series early
23/02/2023	6.9	73.229, 38.055	Tajikistan	Solve for coseismic offset
25/10/2018	6.8	20.557, 37.520	Greece	Solve for coseismic offset
24/01/2020	6.7	39.061, 38.431	Elazığ, Türkiye	Solve for coseismic offset
30/10/2016	6.6	13.096, 42.862	Norcia, Italy	Started time-series late
25/11/2016	6.6	73.978, 39.273	Pamir	Started time-series late
07/01/2022	6.6	101.29, 37.828	China	Solve for coseismic offset
05/09/2022	6.6	102.236, 29.679	China	Solve for coseismic offset
08/08/2017	6.5	103.855, 33.193	China	Solve for coseismic offset
26/06/2016	6.4	73.339, 39.479	Pamir	Started time-series late
17/11/2017	6.4	94.984, 29.8333	China	Solve for coseismic offset
29/12/2020	6.4	16.257, 45.424	Croatia	Solve for coseismic offset
26/11/2019	6.4	19.526, 41.514	Albania	Solve for coseismic offset
14/11/2021	6.4	56.072, 27.727	Iran	Solve for coseismic offset
25/11/2018	6.3	45.744, 34.361	Iran-Iraq	Solve for coseismic offset
08/08/2017	6.3	82.832, 44.302	Tian Shan	Solve for coseismic offset
22/07/2020	6.3	86.864, 33.143	Tibetan Plateau	Solve for coseismic offset
07/10/2023	6.3	61.926, 34.5982	Herat, Afghanistan	Ended time-series early
25/06/2020	6.3	82.416, 35.595	Tibetan Plateau	Solve for coseismic offset
03/03/2021	6.3	22.176, 39.755	Greece	Solve for coseismic offset
05/04/2017	6.1	60.436, 35.776	Iran	Solve for coseismic offset
01/12/2017	6.1	57.307, 30.746	Iran	Solve for coseismic offset
21/05/2021	6.1	100.008, 25.727	China	Solve for coseismic offset
23/11/2022	6.1	30.983, 40.836	Türkiye	Solve for coseismic offset
19/01/2020	6.0	77.108, 39.835	China	Solve for coseismic offset

Table SA3: List of earthquakes with potential impact on the velocity field and strain rates, and the approach used to mitigate the effect. Earthquake moment magnitudes and epicentral locations from the USGS are given. Full information on the impacted frames is given in Supplementary Material. Note that any major aftershocks of these events will also be accounted for in the mitigation process when truncating the time series. Locations of earthquakes and moment tensors are shown in Figure SB12.

Appendix B. Supplementary Appendix Figures

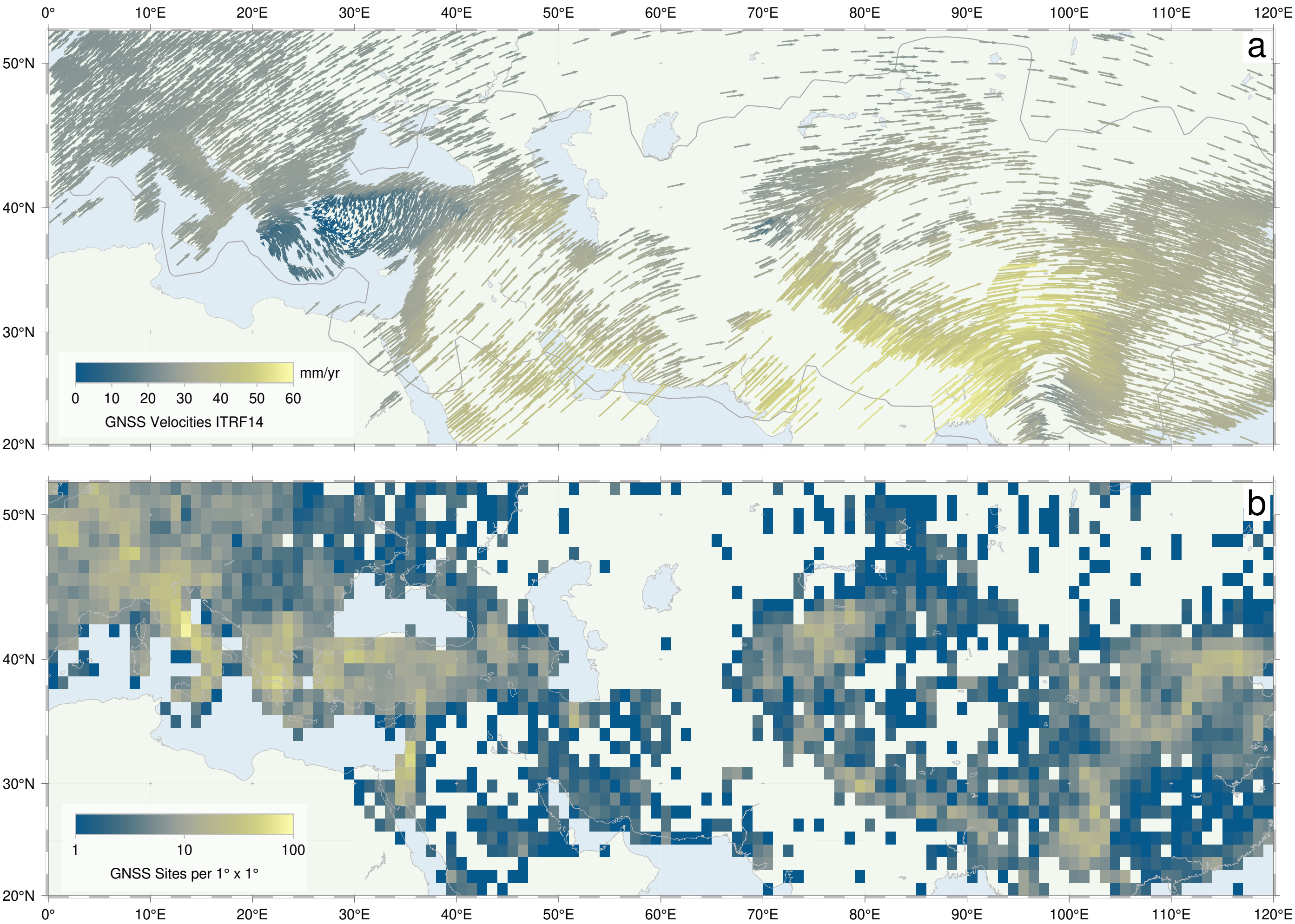

Figure SB1: (a) Compiled and aligned GNSS velocities with respect to ITRF14. The dark grey polygon denotes the coverage area for the combined InSAR-GNSS velocity field. (b) Number of GNSS sites per one degree cell (note logarithmic scale).

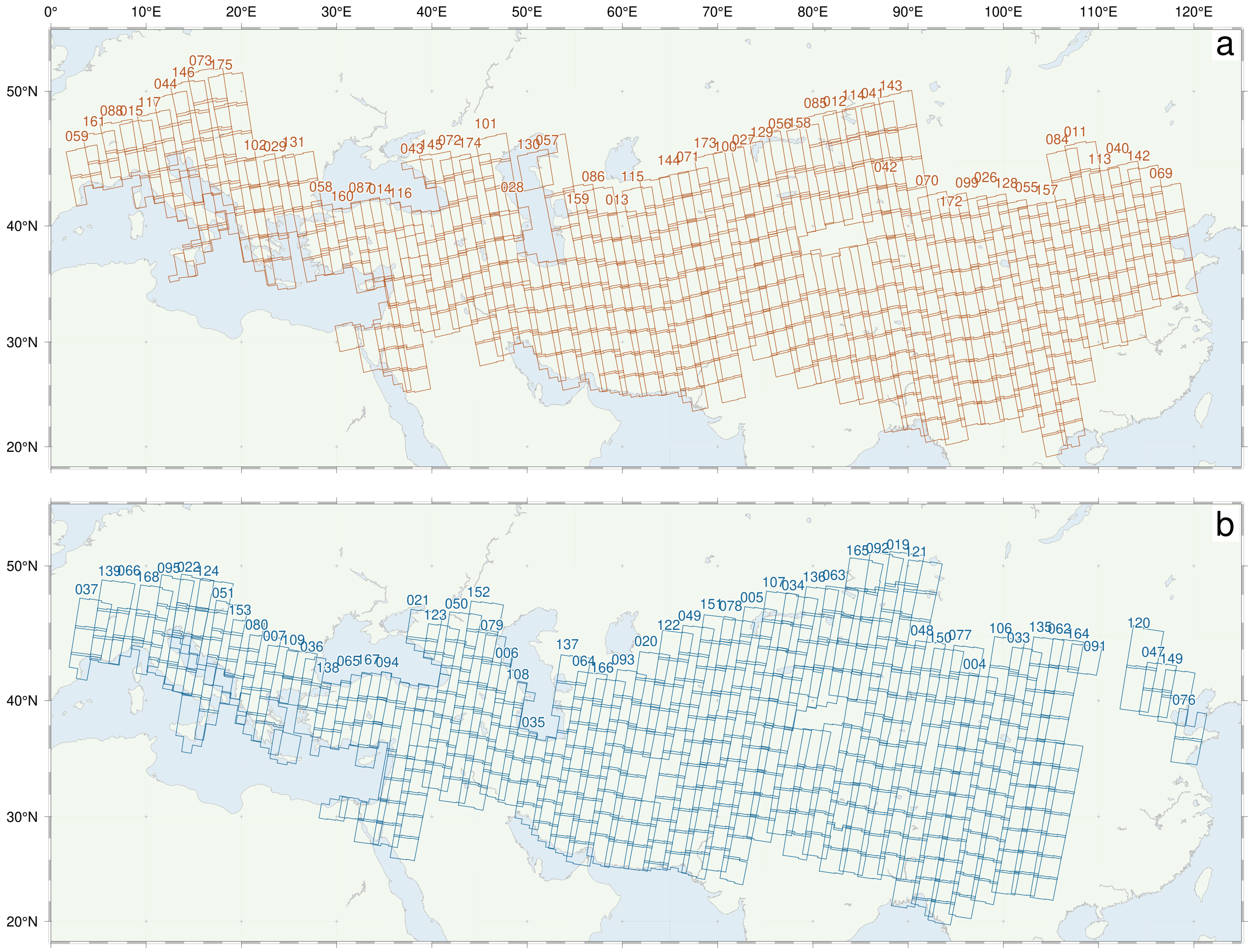

Figure SB2: Sentinel-1 data coverage from frames defined within the LiCSAR system on (a) ascending (brown) and (b) descending (blue) track directions. Each frame typically consists of 13 bursts on each of the 3 subswaths with one burst overlap with adjacent along track frames. Frames near coasts or at the edges of the area of interests will have differing number of bursts. The dark grey polygon denotes the coverage area for the combined InSAR-GNSS velocity field mesh.

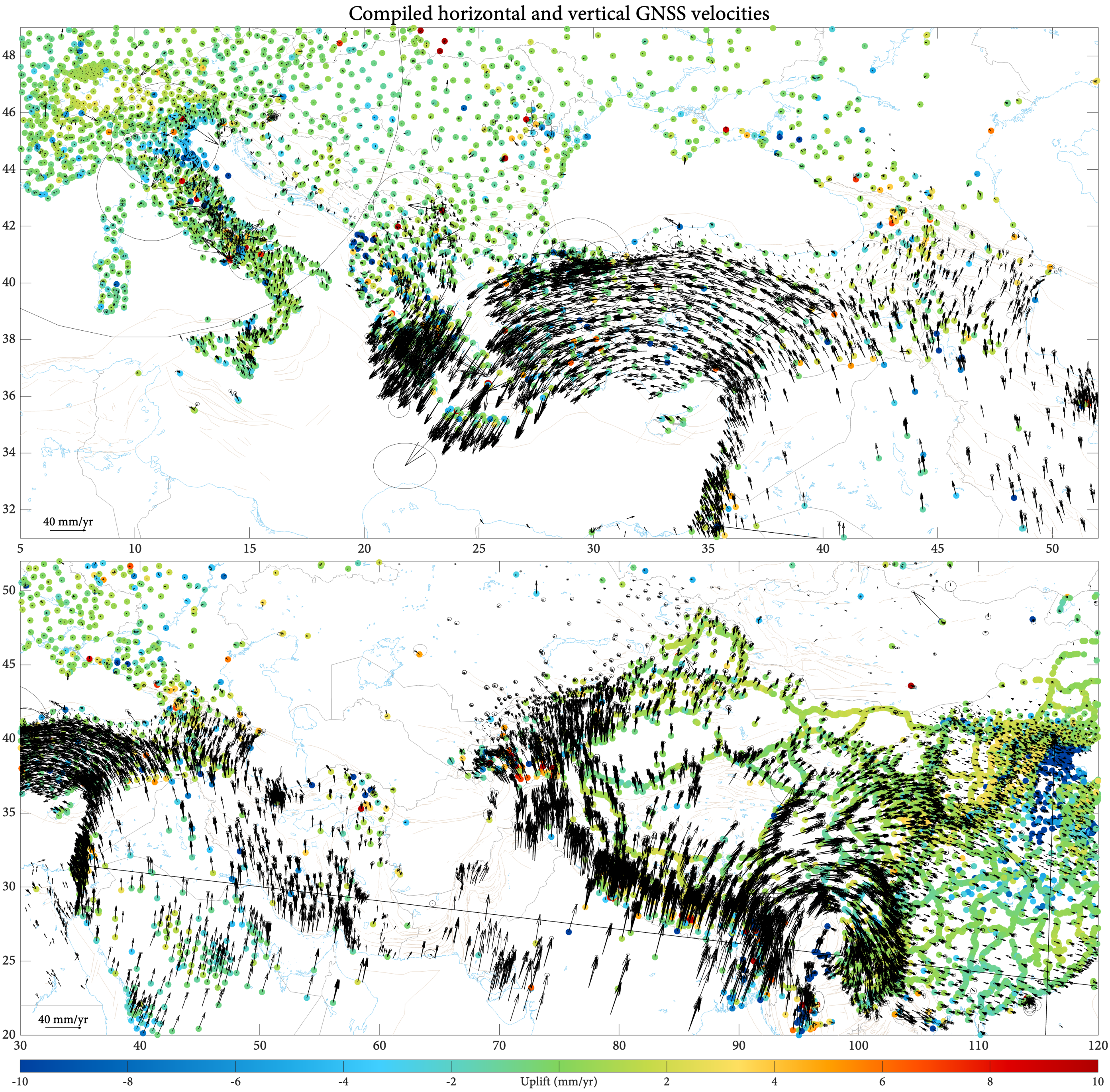

Figure SB3: The GNSS and other point velocities compiled from the published literature (Table SA1), in the western AHB (top) and eastern AHB (bottom). Arrows show horizontal motions with respect to the ITRF2014 Eurasia-fixed reference frame (Altamimi et al., 2017). Coloured circles denote the uplift rate for sites with a vertical component.

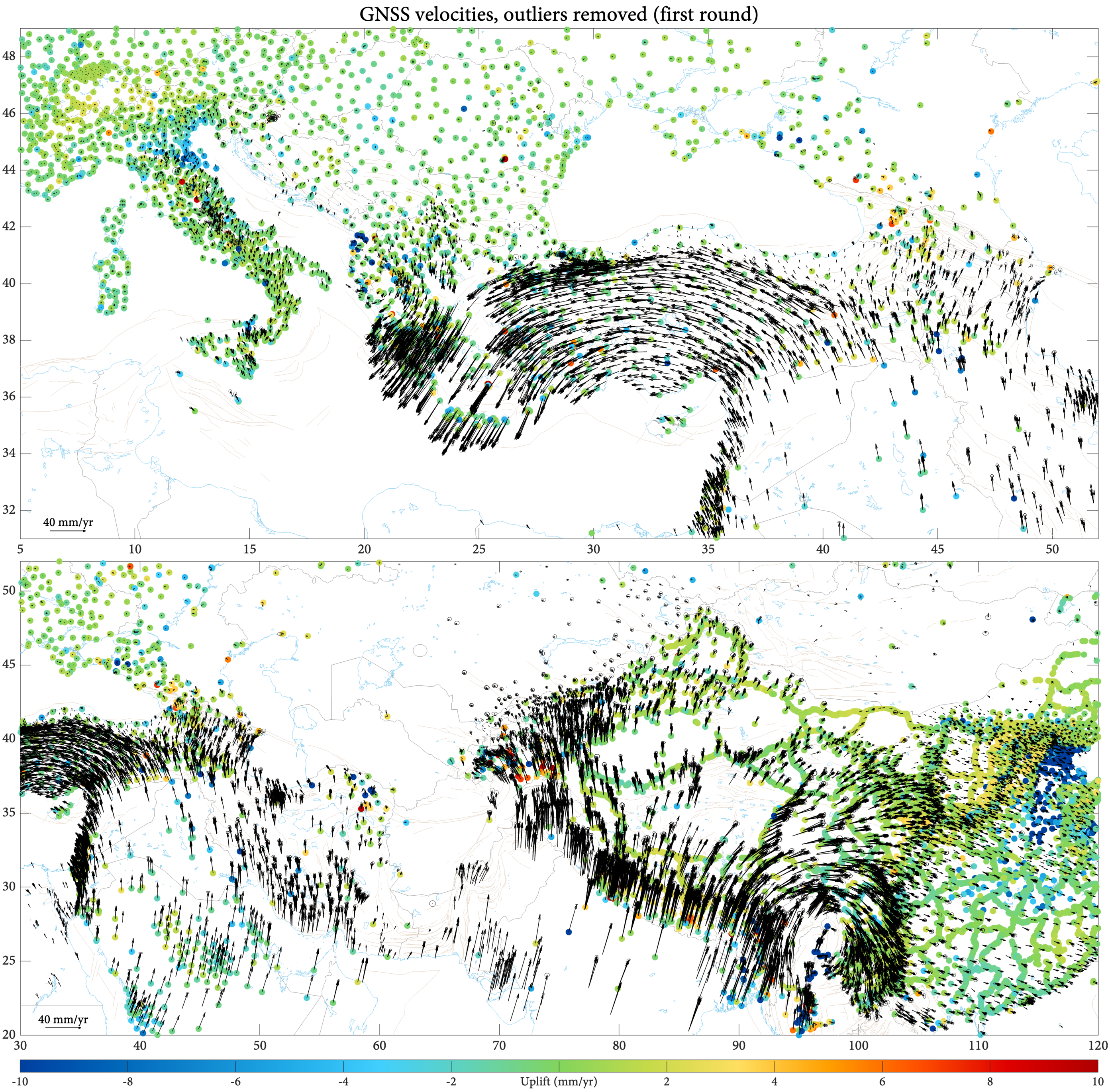

Figure SB4: GNSS and other point velocities after the first round of removal of outliers.

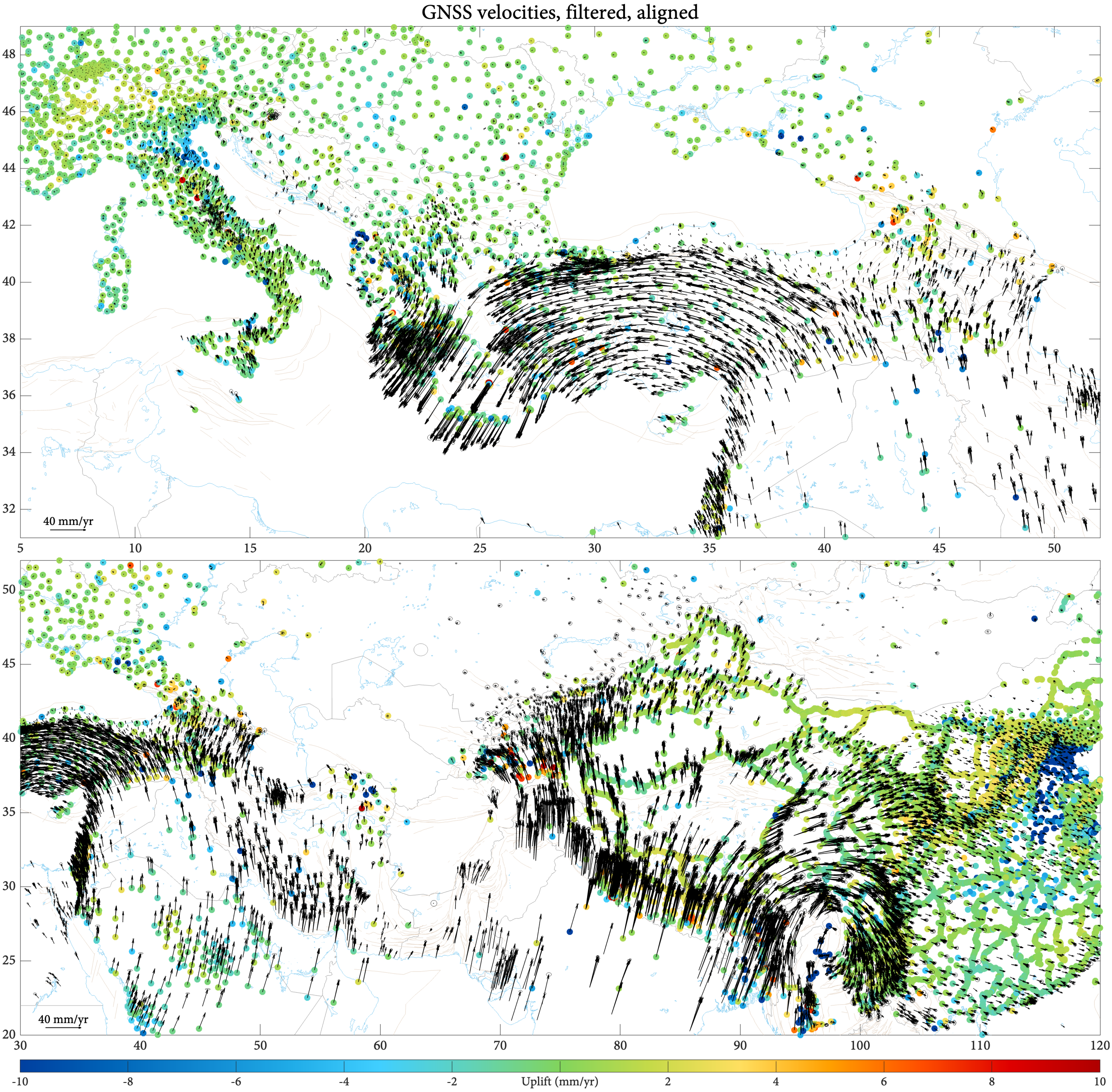

Figure SB5: GNSS and other point velocities after the alignment step (iterative removal of systematic differences via best-fit Euler-pole rotations).

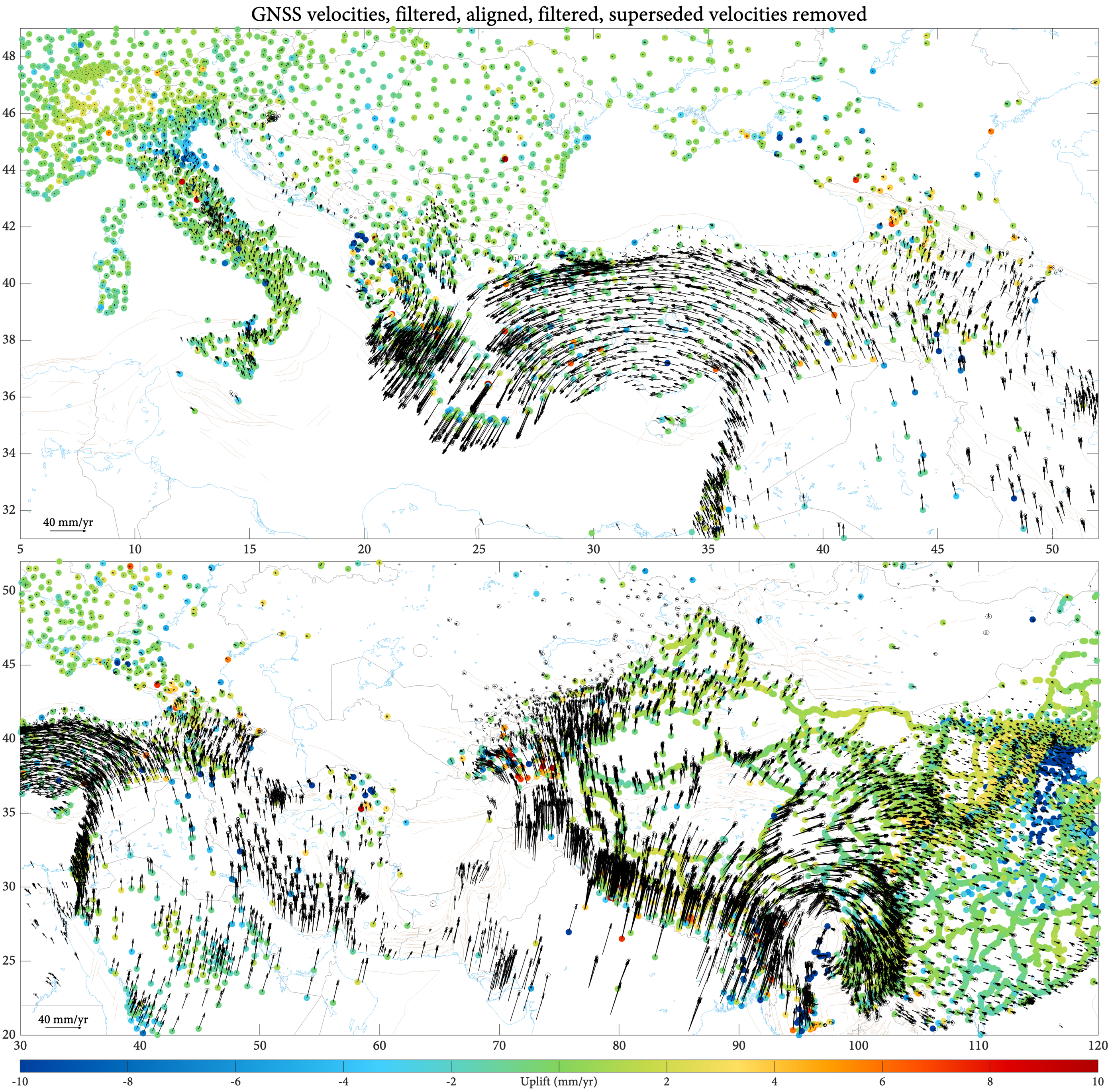

Figure SB6: GNSS and other point velocities after the second round of outlier removal and the removal of superseded and duplicate velocities.

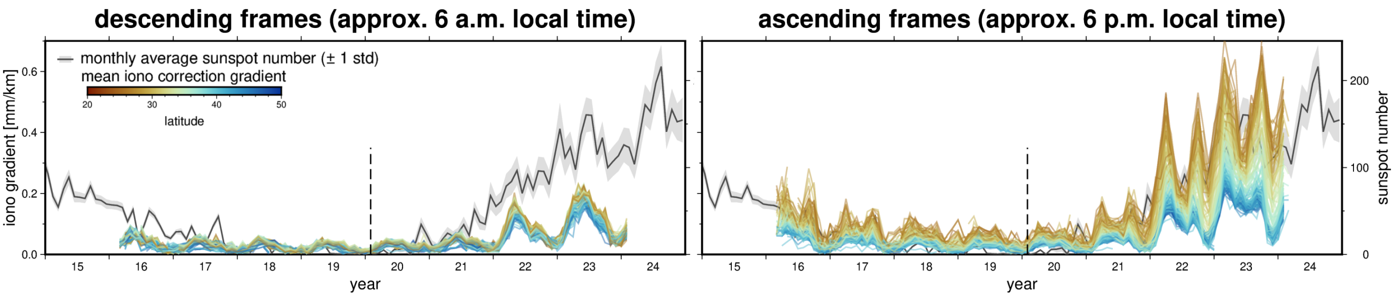

Figure SB7: Ionosphere gradient corrections for the descending (left) and ascending (right) frames in their monthly averages, plotted with the timeline of sunspot numbers through cycle 24 and 25 (2014–2024). Sunspot data from the World Data Center SILSO, Royal Observatory of Belgium, Brussels (Clette and Lefèvre, 2015). The correction gradients are calculated as mean absolute gradients in the direction of maximum gradient per temporal epoch w.r.t. nearest epoch to the 2020-02-01 (vertical dashed line).

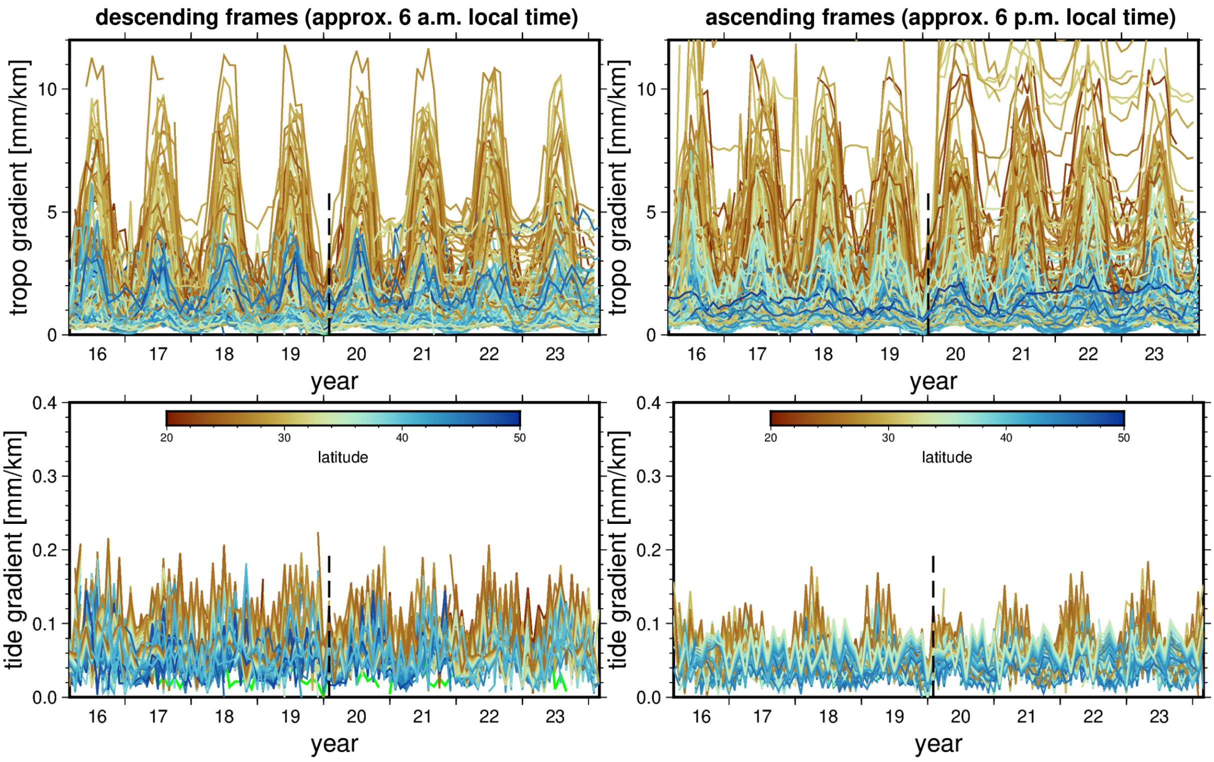

Figure SB8: Troposphere (top) and solid earth tide (bottom) gradient corrections for the descending (left) and ascending (right) frames. The correction gradients are calculated as mean absolute gradients in the direction of maximum gradient per temporal epoch w.r.t. nearest epoch to the 2020-02-01 (vertical dashed line).

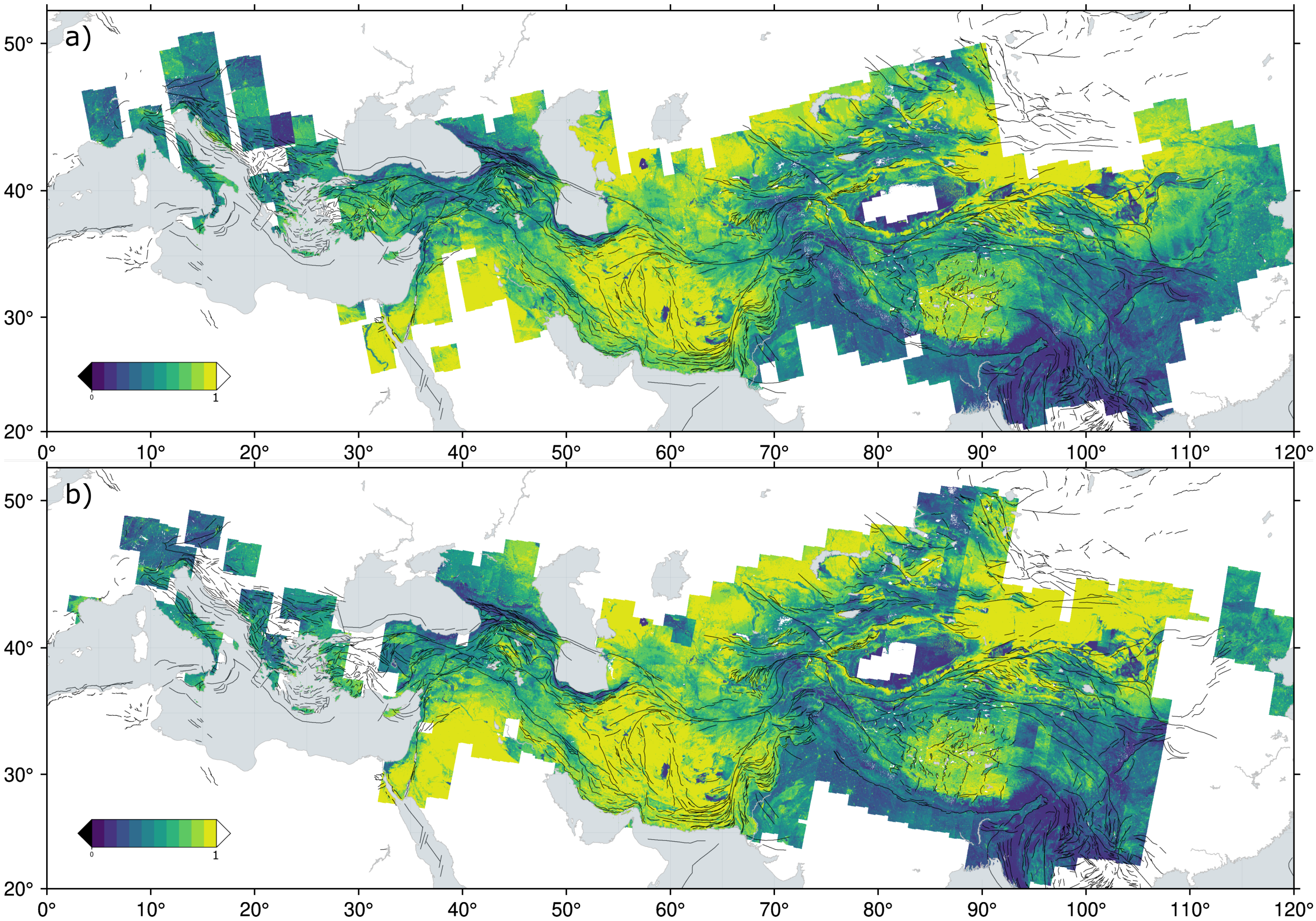

Figure SB9: Average coherence calculated for frames in a) ascending and b) descending directions.

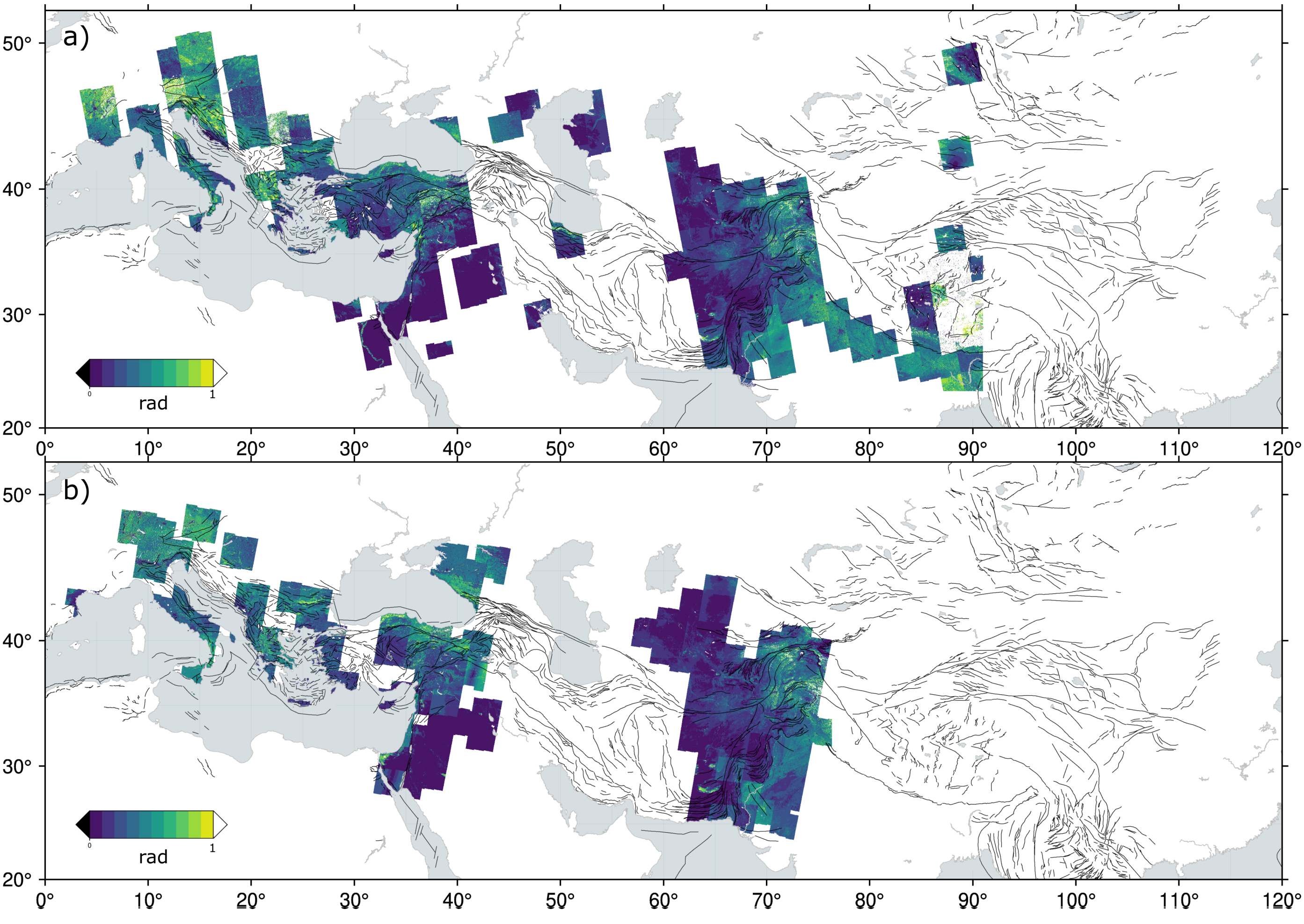

Figure SB10: Average absolute phase bias (rad) as calculated for frames in a) ascending and b) descending directions.

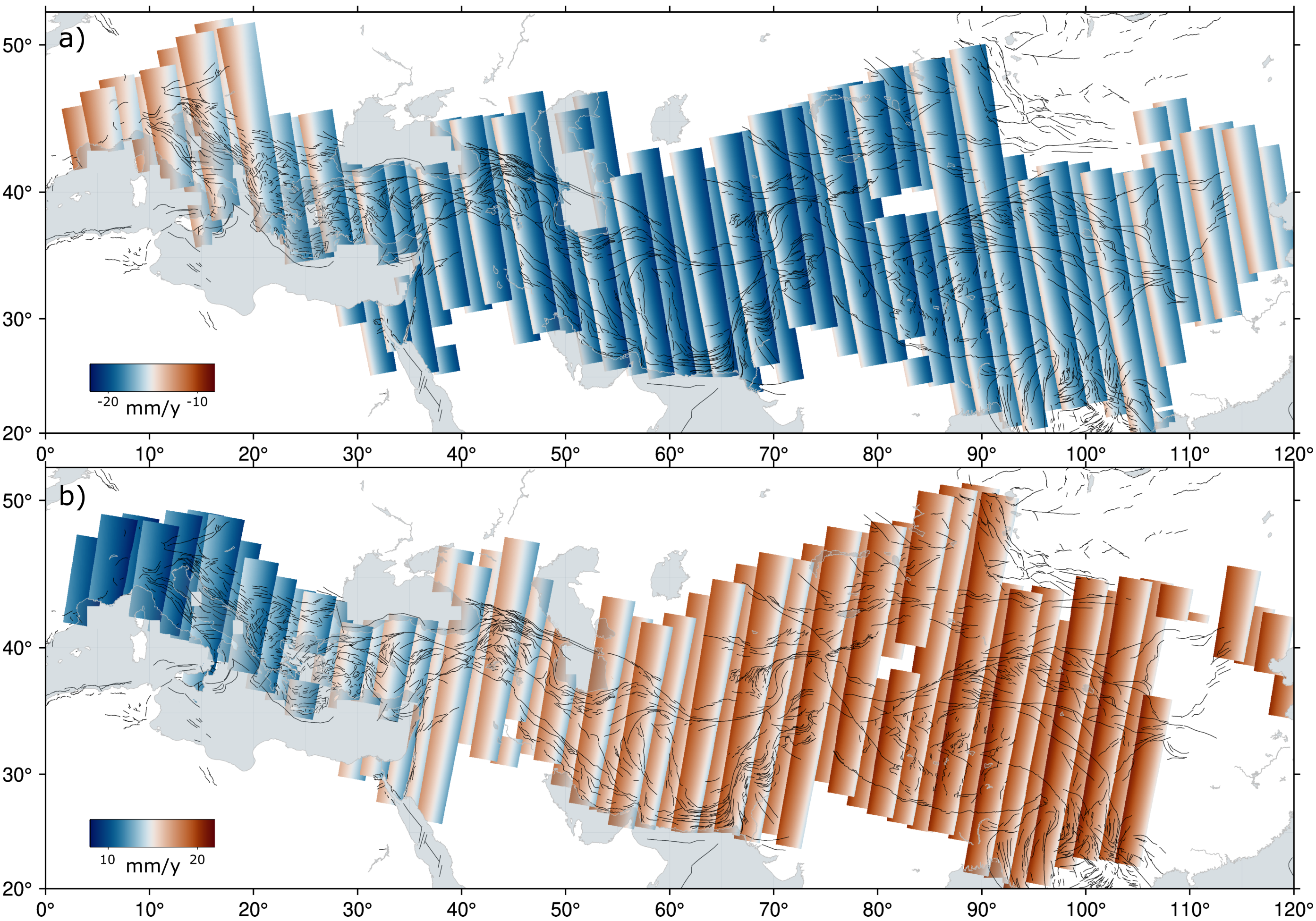

Figure SB11: Plate motion correction to the line of sight average velocities (mm/yr) for the tracks in the a) ascending and b) descending directions.

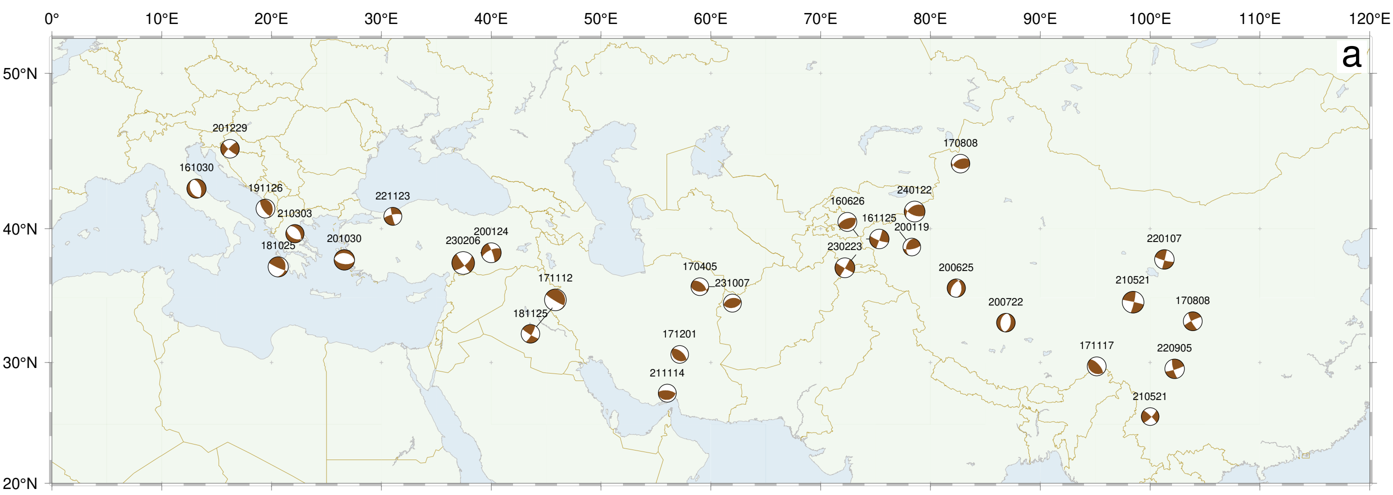

Figure SB12: Location of M_w > 6.0 earthquakes (denote by focal mechanism from the GCMT, (Ekström et al., 2012)) that are accounted for in the time series (Table SA3). Numbers above the focal mechanism denote the date of the earthquake (UTC, YYMMDD).

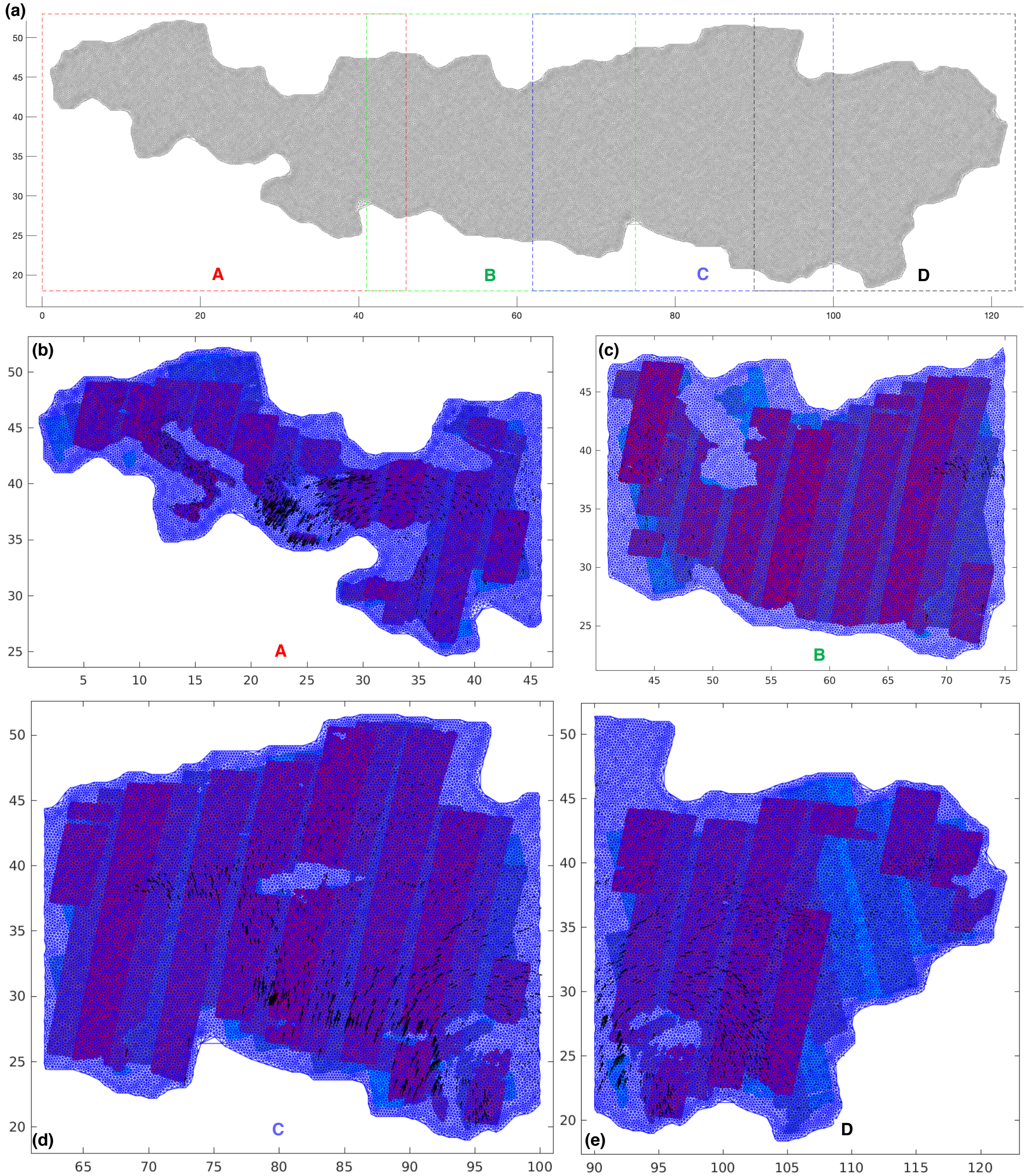

Figure SB13: (a) Triangular mesh constructed for the entire AHB, used for VELMAP processing. The mesh is divided into four regions (A, B, C, and D), marked by dashed rectangles. Panels (b–e) show zoomed-in views of these regions, corresponding to four separate VELMAP runs.

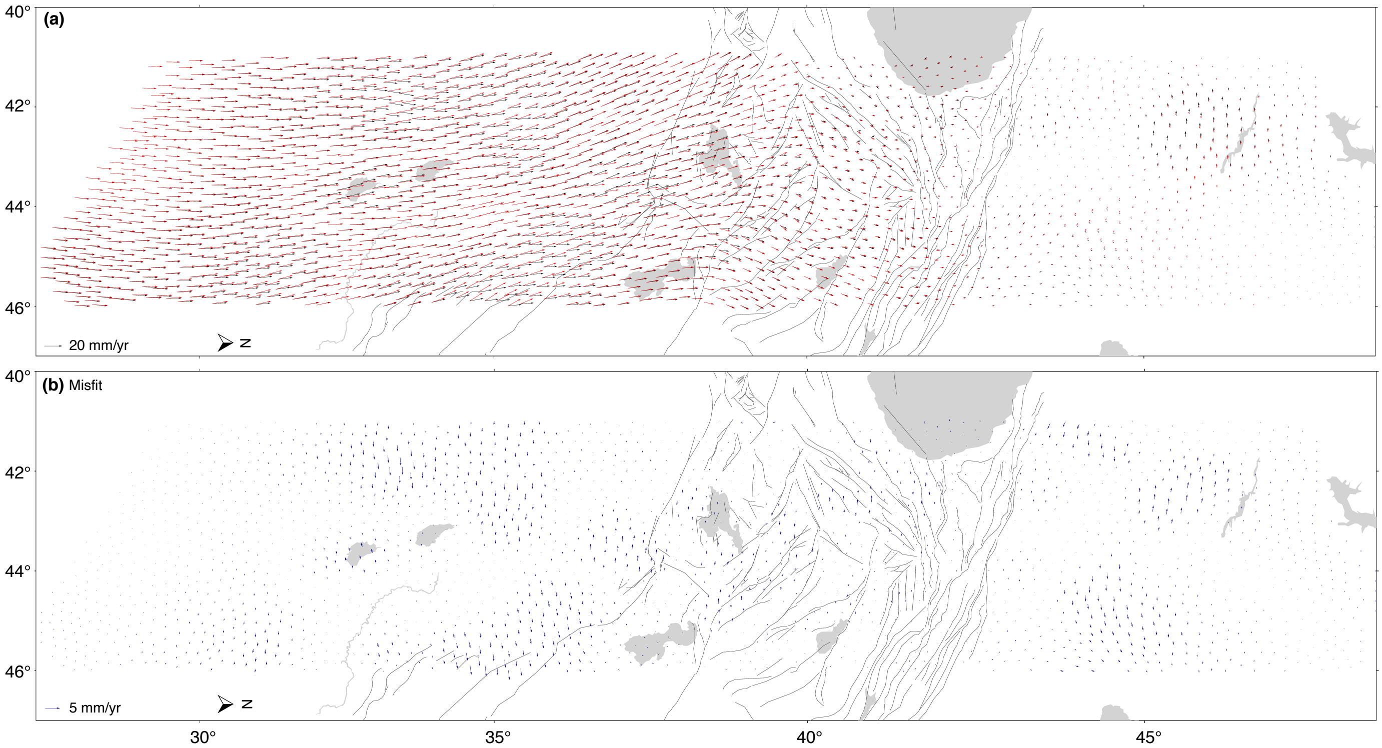

Figure SB14: (a) Velocity solutions at the common nodes in the overlapping region between A and B from separate VELMAP runs, as shown in Figure SB13. (b) Mismatch in the overlap. Note that the figure is rotated 90° clockwise for better visualization.

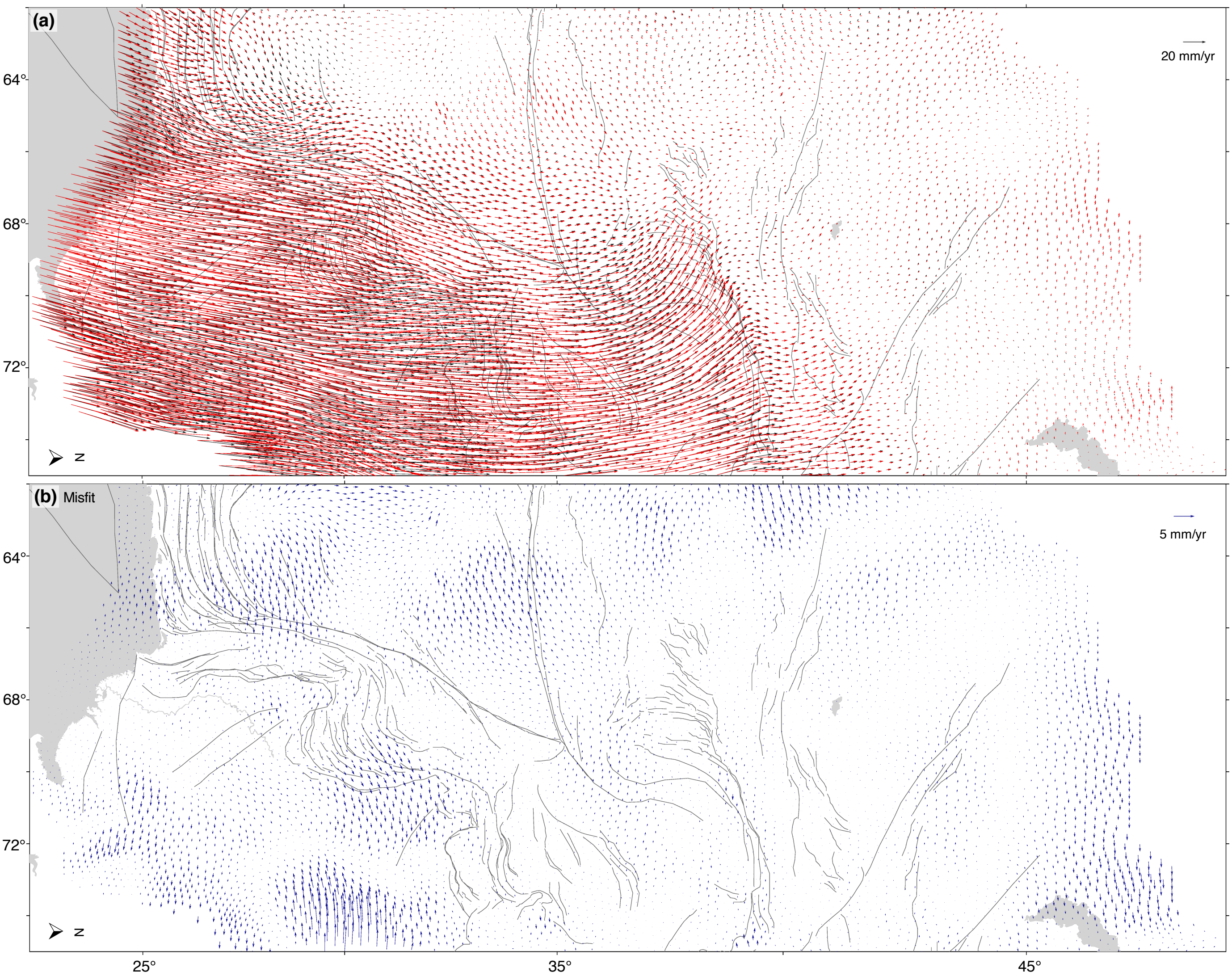

Figure SB15: (a) Velocity solutions at the common nodes in the overlapping region between B and C from separate VELMAP runs, as shown in Figure SB13. (b) Mismatch in the overlap. Note that the figure is rotated 90° clockwise for better visualization.

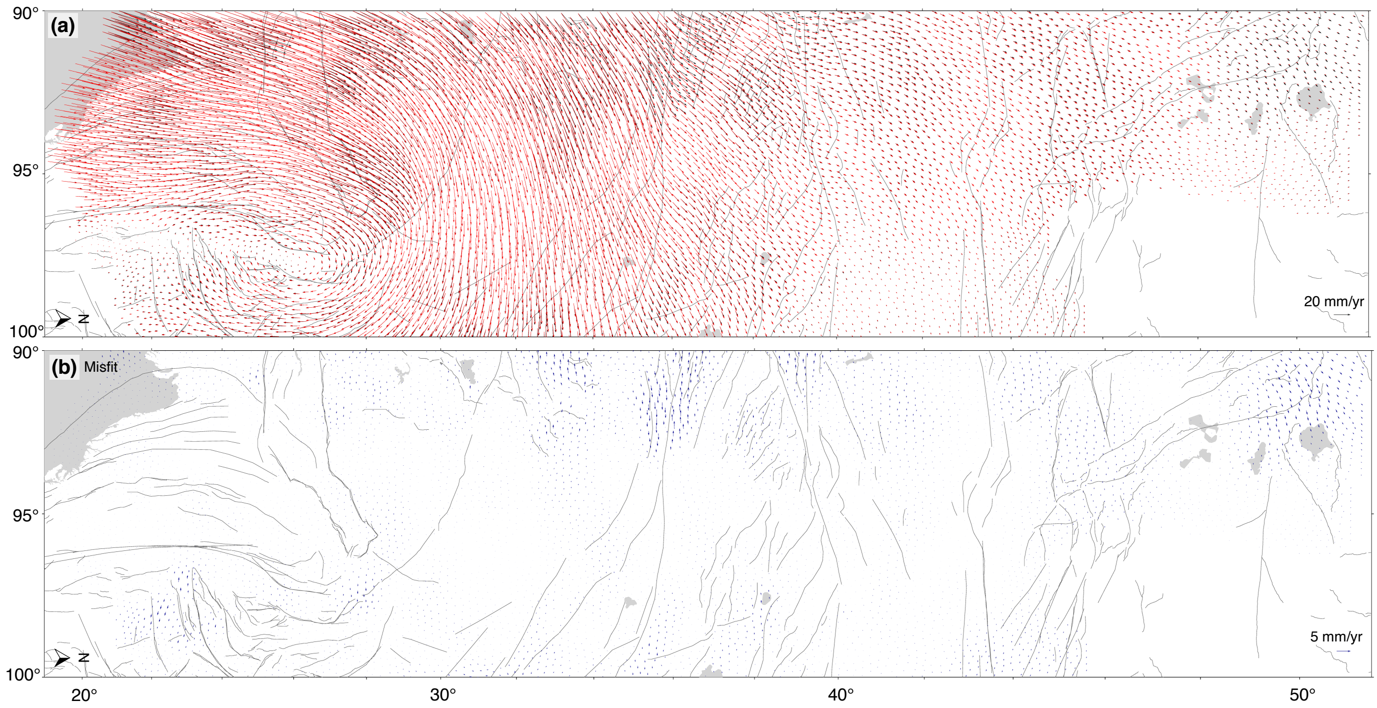

Figure SB16: (a) Velocity solutions at the common nodes in the overlapping region between C and D from separate VELMAP runs, as shown in Figure SB13. (b) Mismatch in the overlap. Note that the figure is rotated 90° clockwise for better visualization.

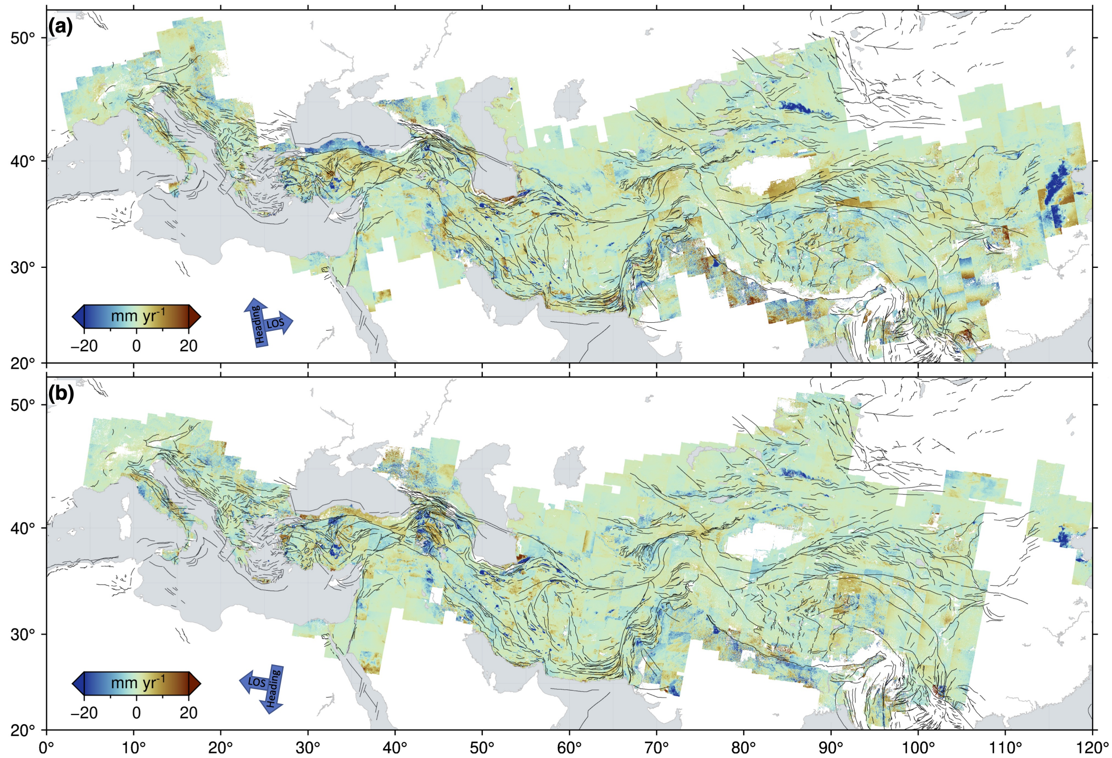

Figure SB17: Individual frame-by-frame line-of-sight velocities on (a) ascending and (b) descending directions, prior to referencing within the tying procedure implemented in VELMAP.

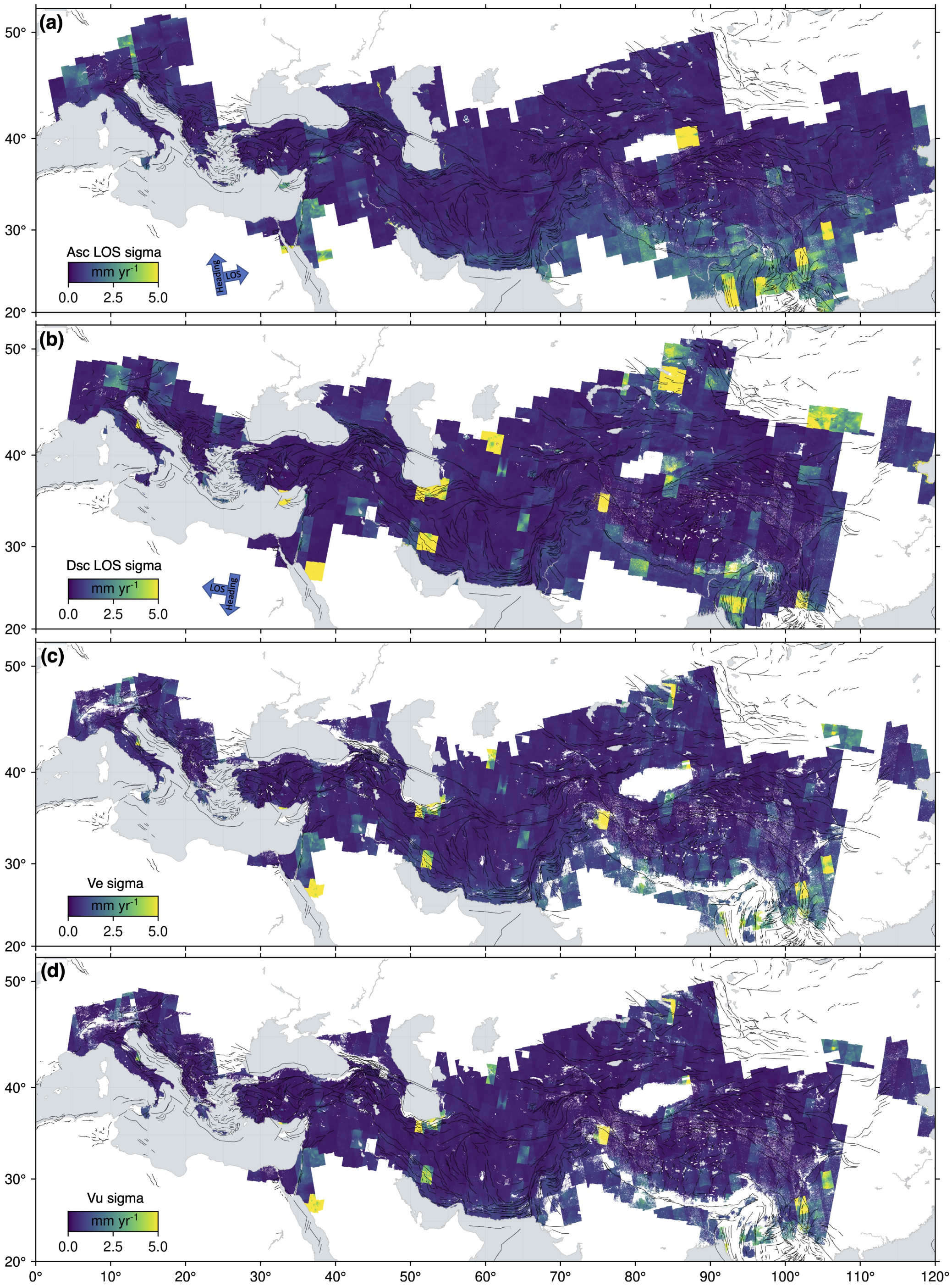

Figure SB18: Uncertainties in line of sight velocities on (a) ascending and (b) descending directions, as well as formal uncertainties of decomposed (c) east and (d) vertical velocities.

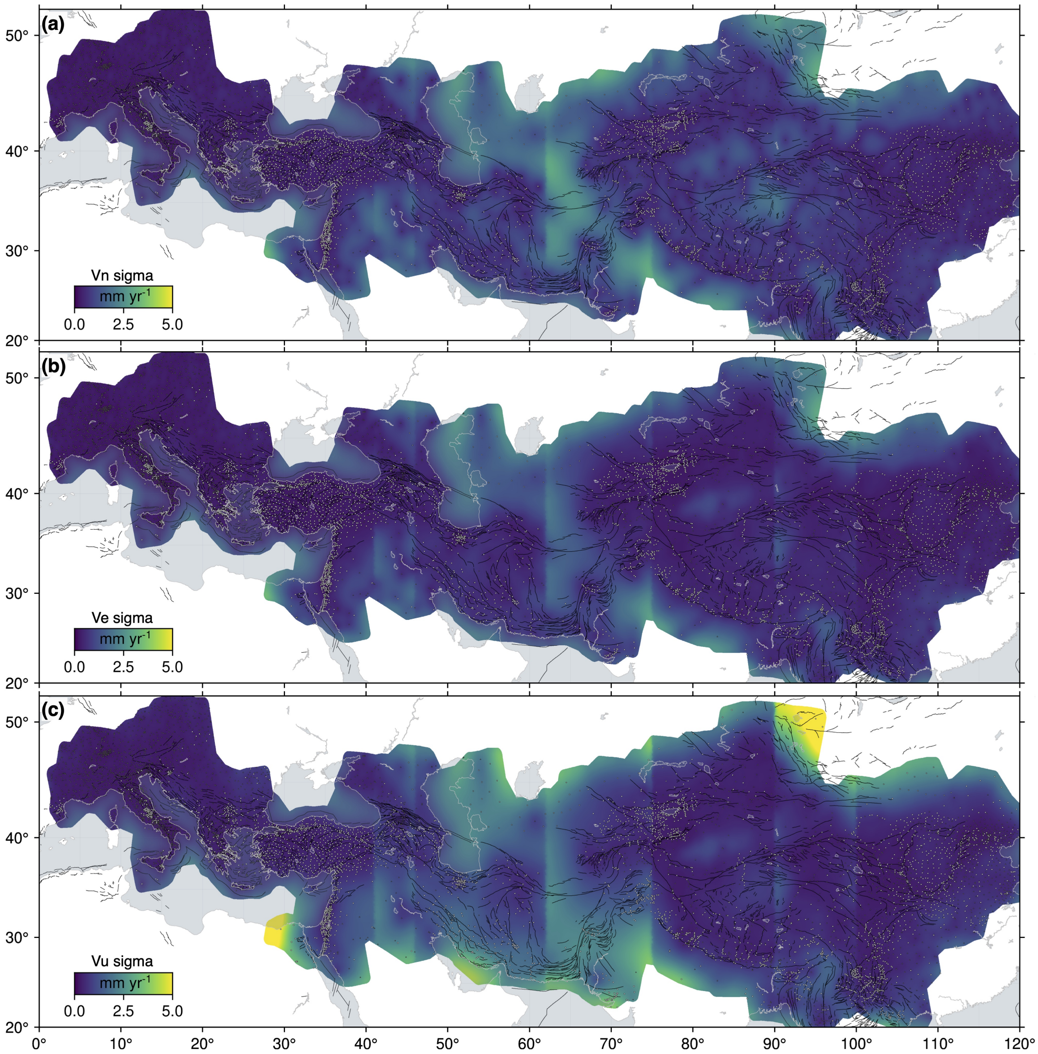

Figure SB19: Uncertainties in VELMAP-derived estimates of (a) north, (b) east, and (c) vertical velocities.

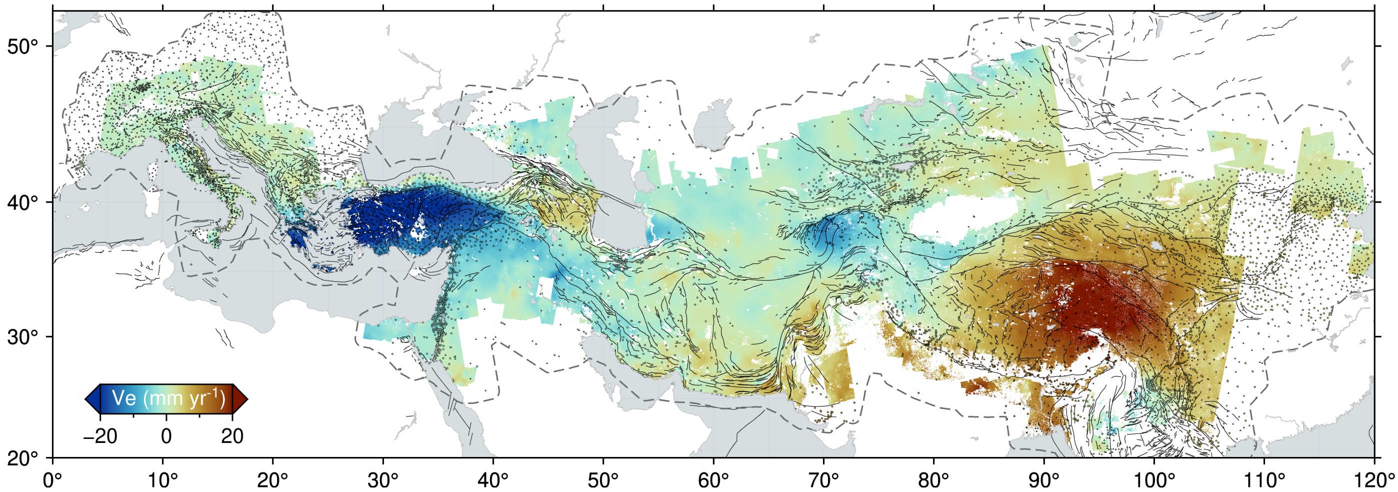

Figure SB20: Gaussian-filtered eastward velocities with a sigma width of 15 km and a window size of 100 km, used for strain rate calculations. Pixels with an absolute vertical velocity magnitude exceeding 20 mm/yr are masked out for strain rate analysis.

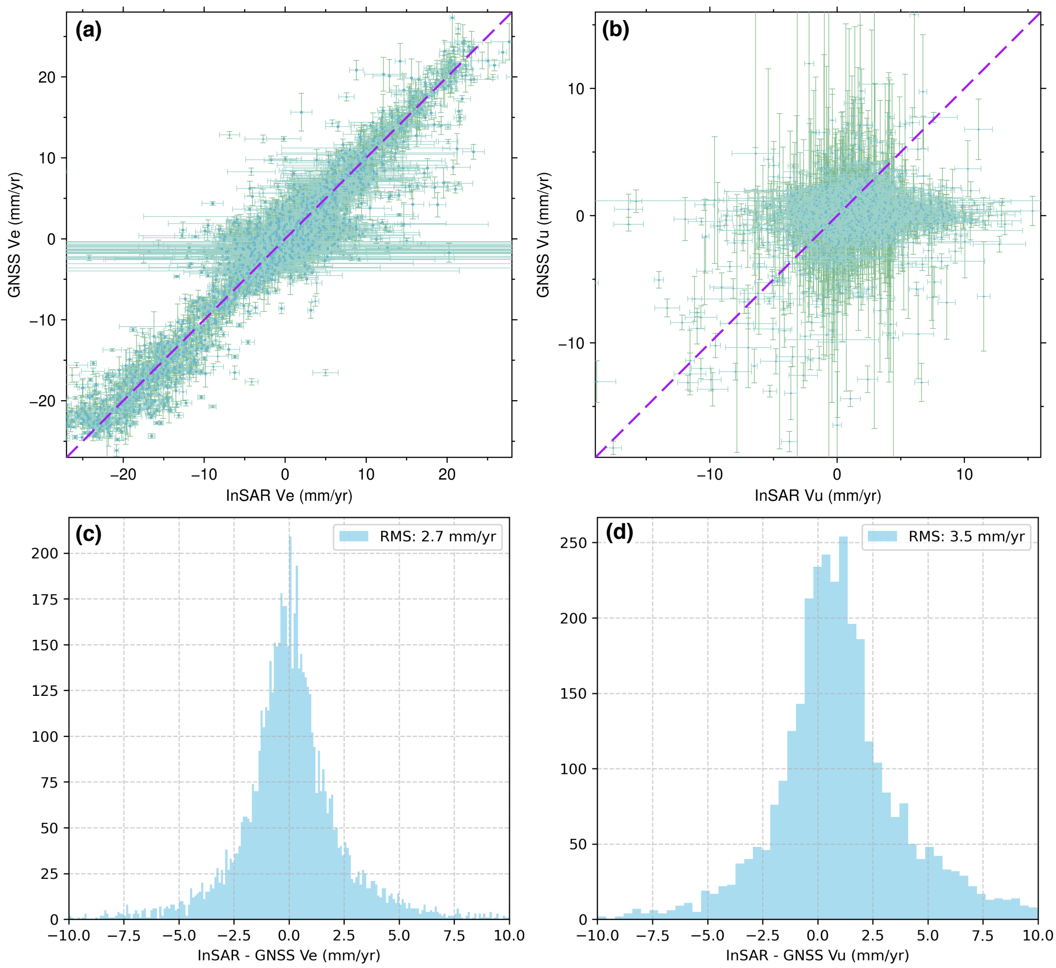

Figure SB21: Correlation plot of InSAR-derived Ve against GNSS Ve across the entire AHB. (b) Correlation plot of InSAR-derived Vu against GNSS Vu. The corresponding histograms of mismatch are shown in panels (c)–(d).

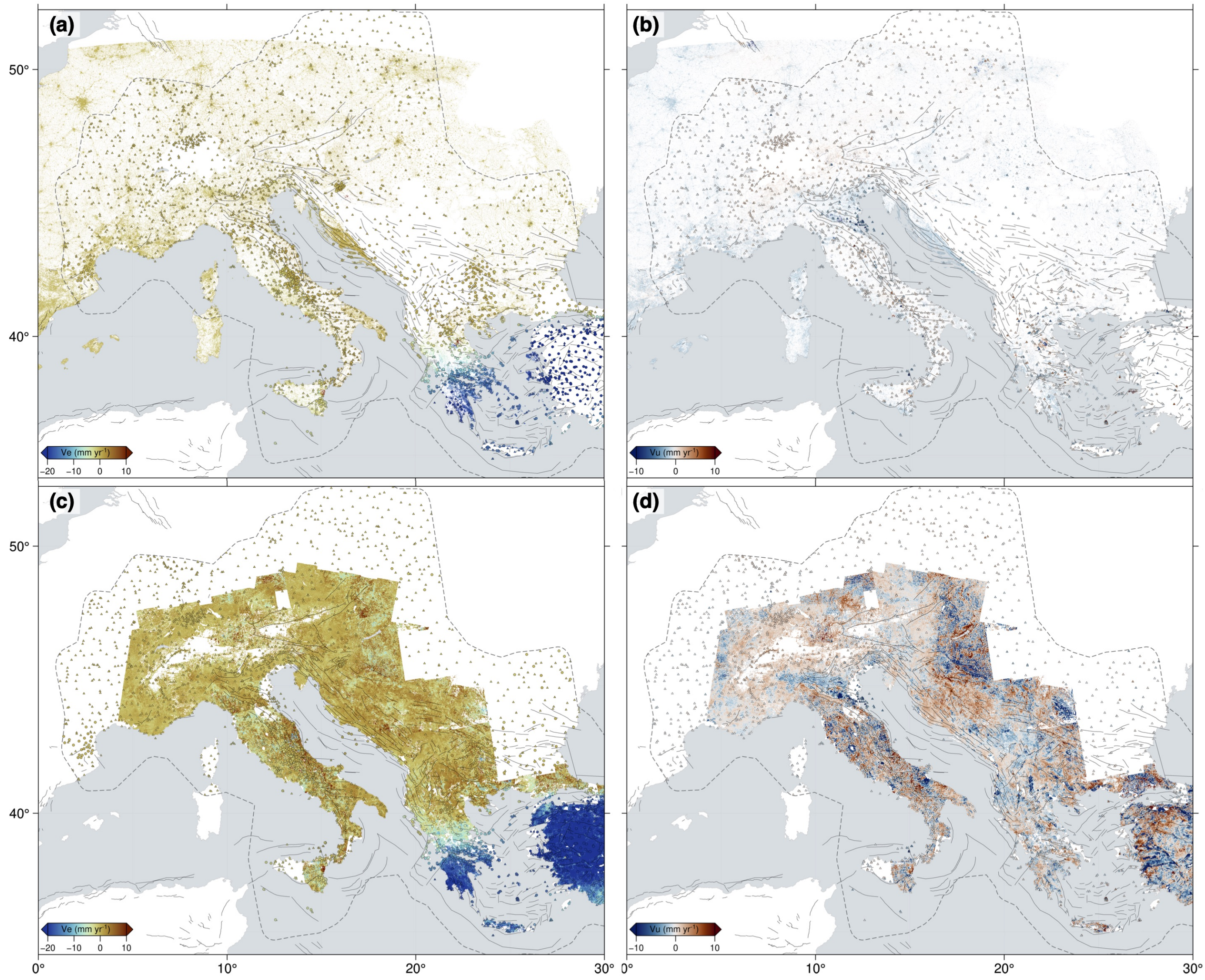

Figure SB22: Comparison with the European Ground Motion Service (EGMS) East and Vertical velocities (Costantini et al., 2021). (a)-(b) EGMS-derived Ve and Vu at a resolution of 100 m. (c)-(d) InSAR-derived Ve and Vu at 1 km resolution. Scatter and histogram plots are shown in Figures SB23 and SB24.

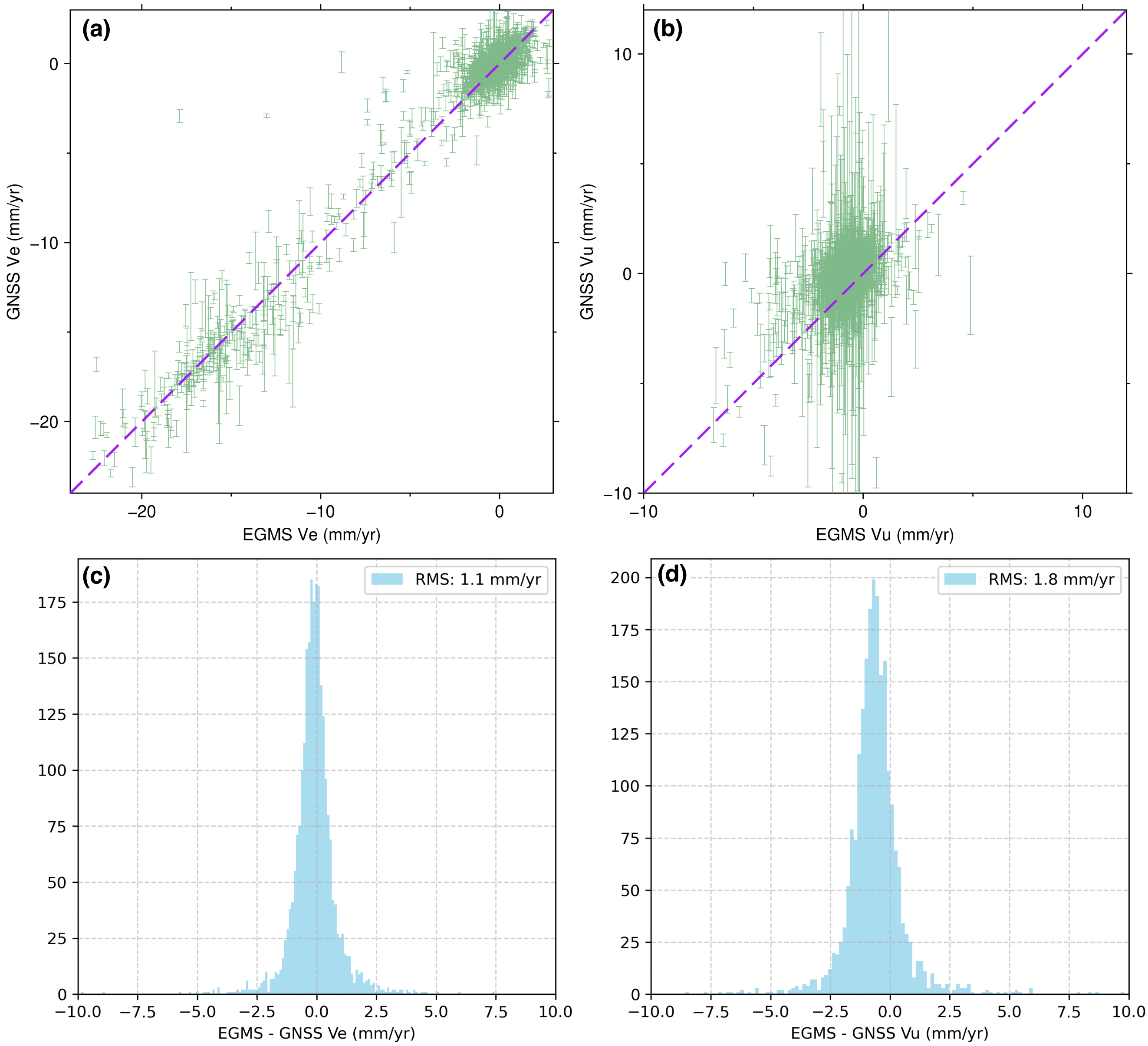

Figure SB23: Scatter and histogram plots comparing EGMS Ve (Vu) with GNSS Ve (Vu).

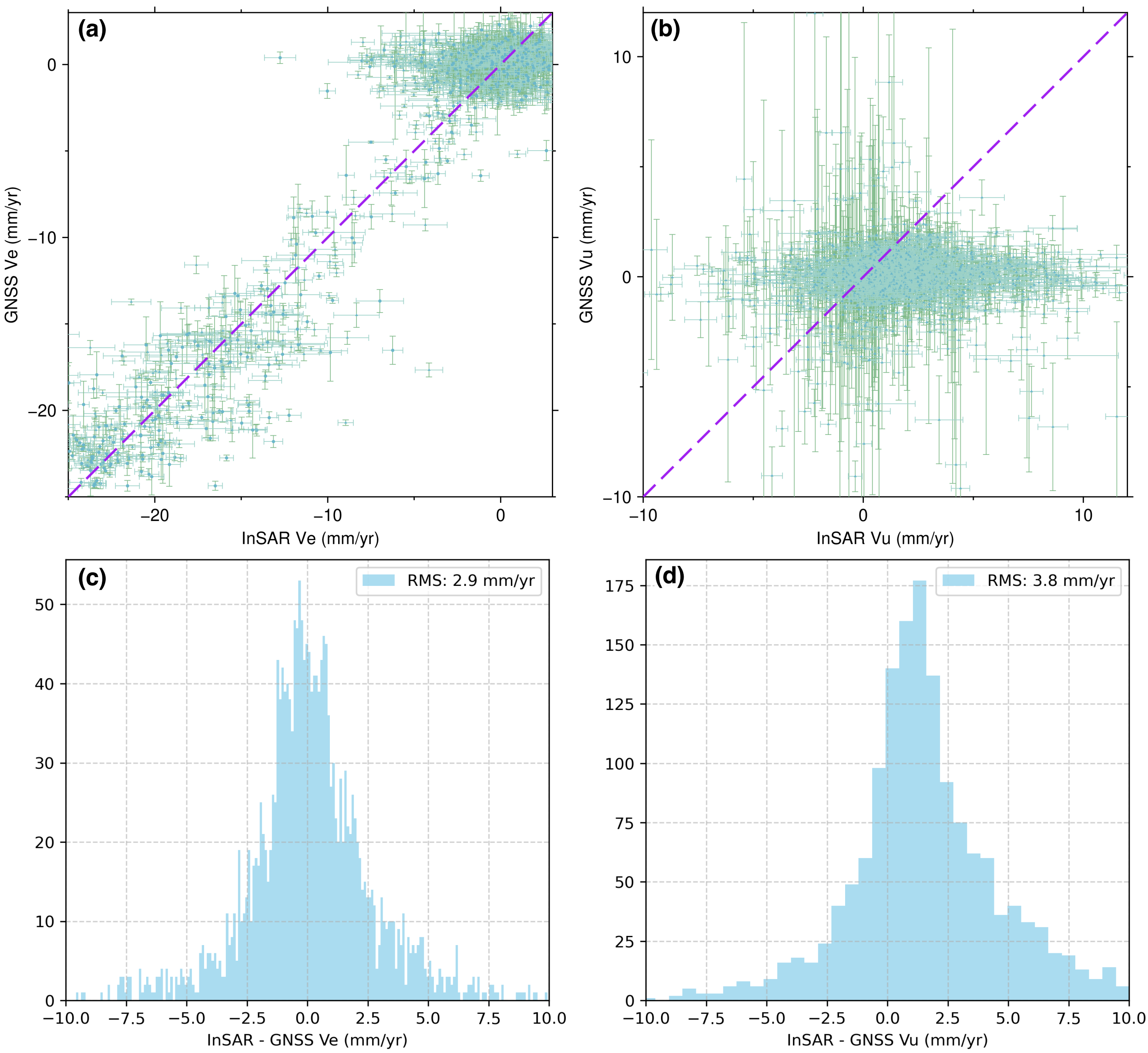

Figure SB24: Scatter and histogram plots comparing InSAR Ve (Vu) with GNSS Ve (Vu) within the same coverage area as the EGMS data.

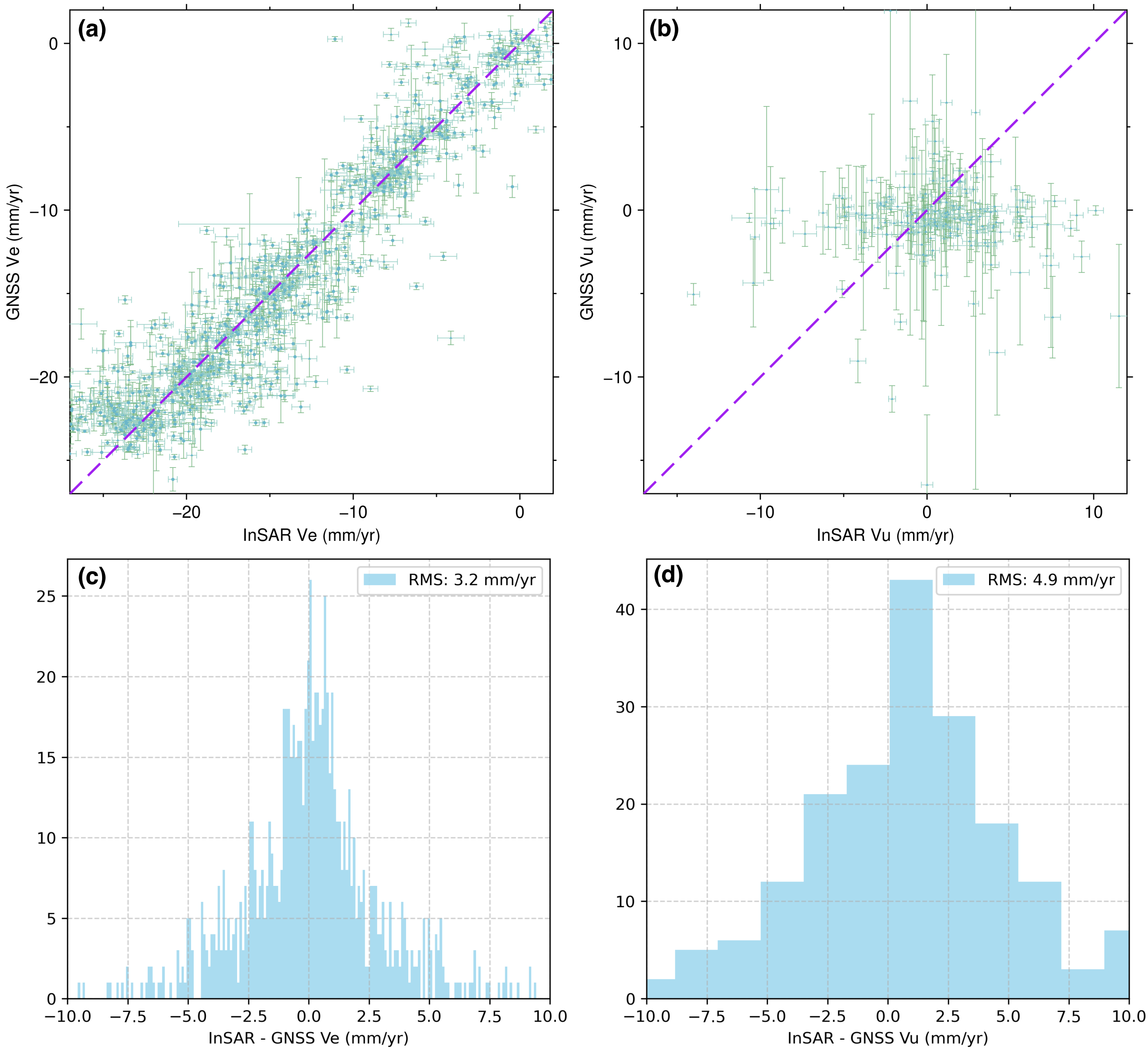

Figure SB25: Scatter and histogram plots comparing InSAR Ve (Vu) with GNSS Ve (Vu) in Anatolia.

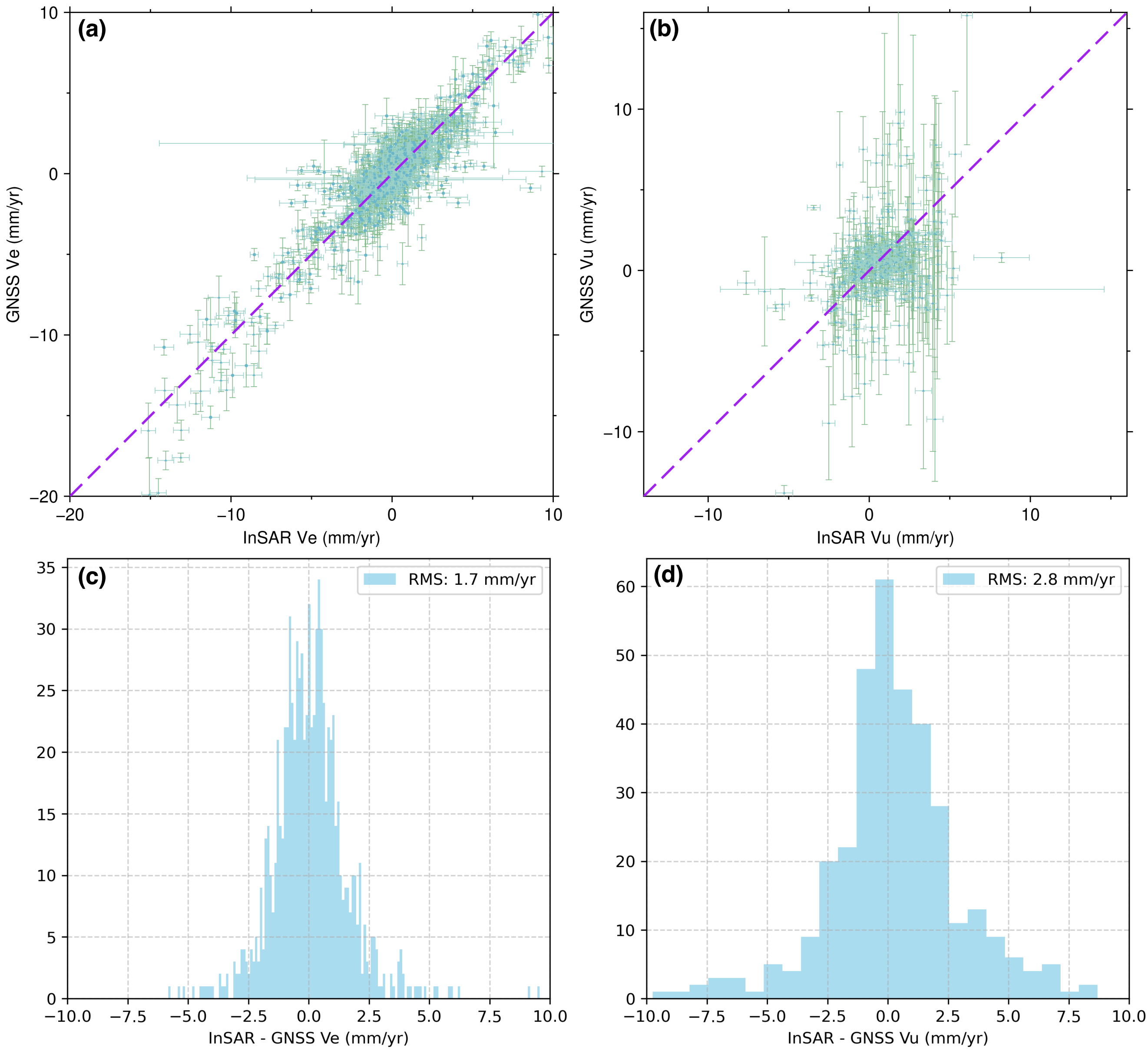

Figure SB26: Scatter and histogram plots comparing InSAR Ve (Vu) with GNSS Ve (Vu) in Pamir and Tian Shan.

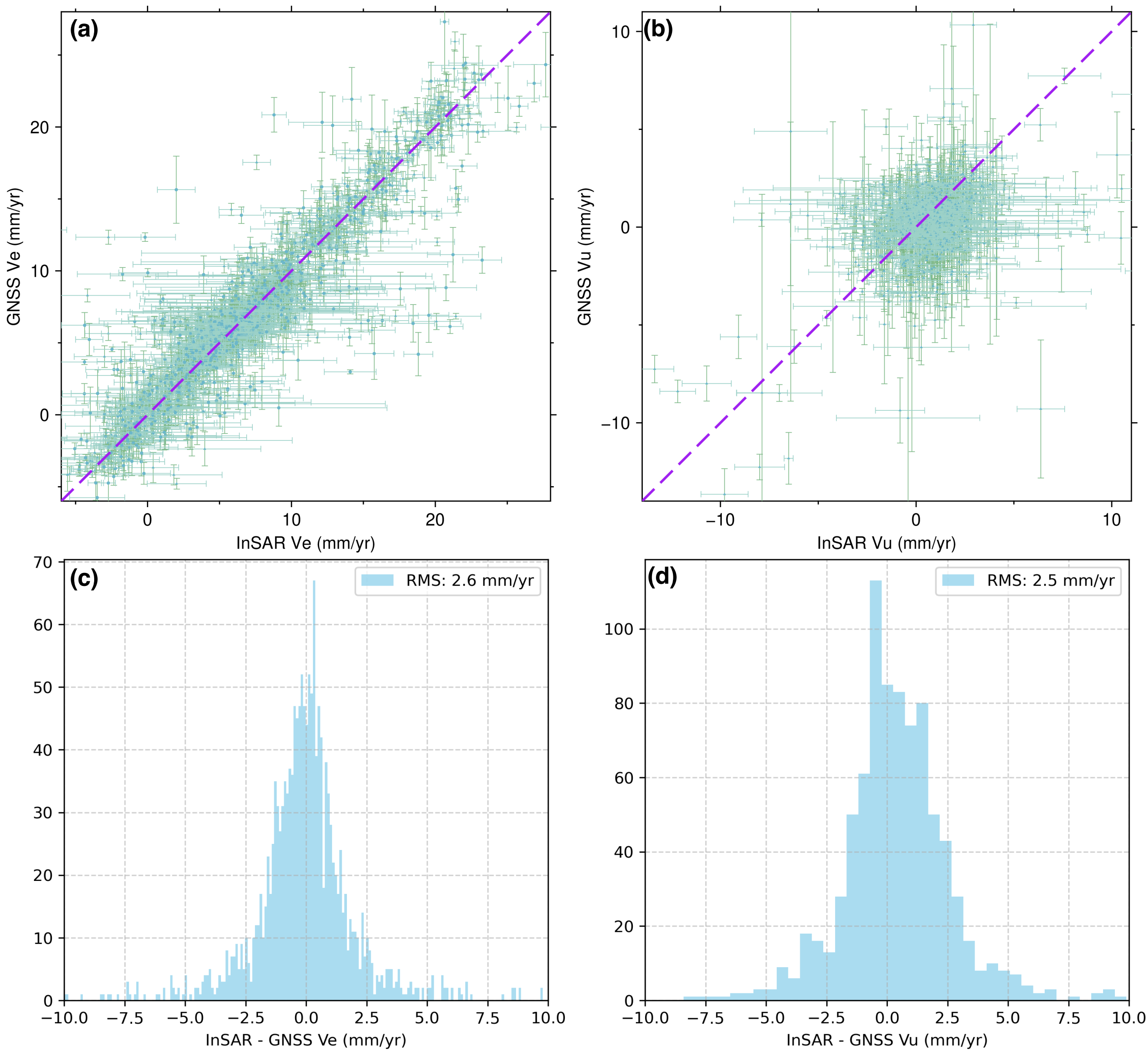

Figure SB27: Scatter and histogram plots comparing InSAR Ve (Vu) with GNSS Ve (Vu) in Tibet.

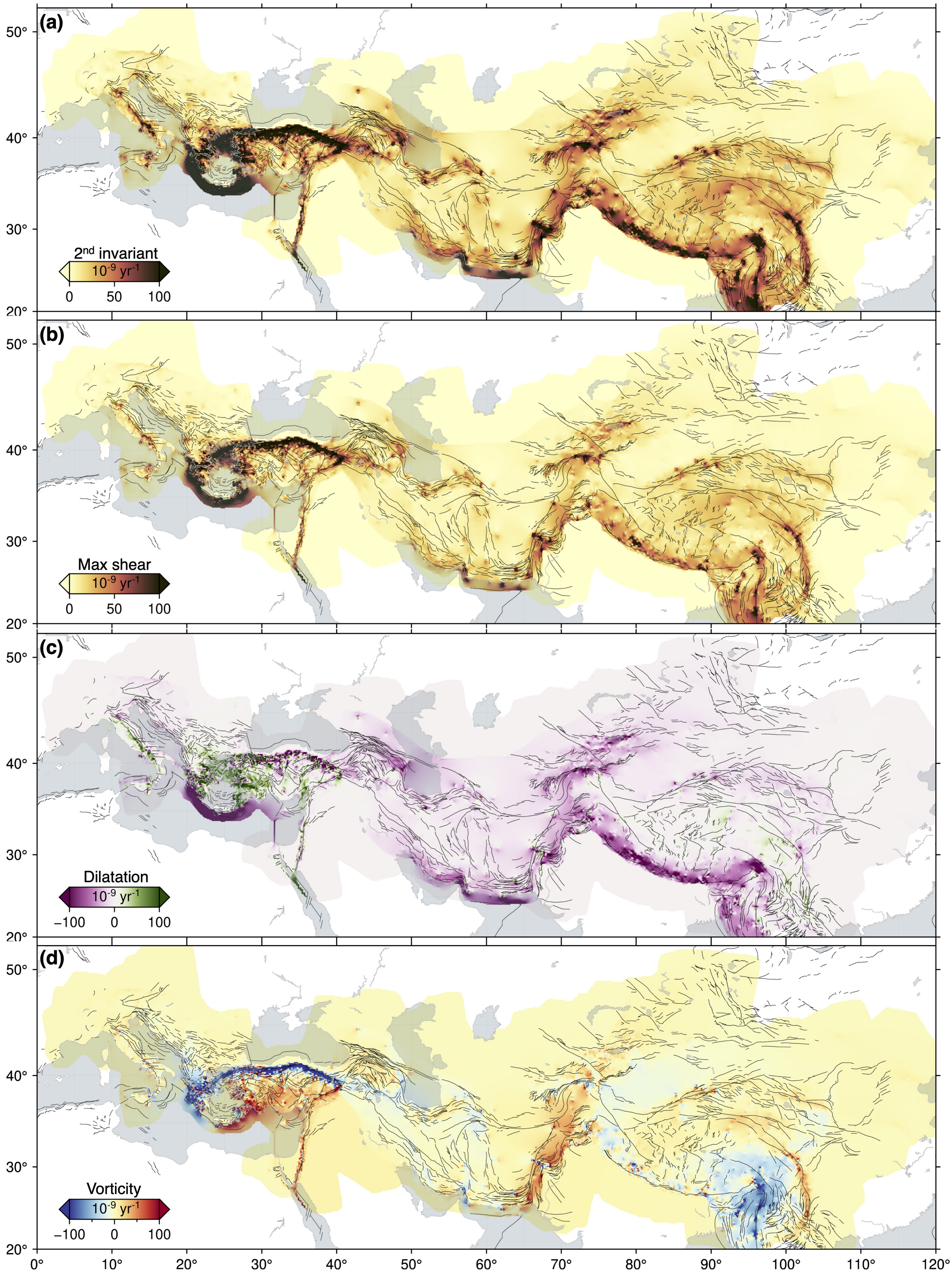

Figure SB28: Global Strain Rate Model (GSRM) v2.1 from Kreemer et al. (2014) (https://geodesy.unr.edu/GSRM/). Compare with Figure 9 for our GNSS+InSAR derived version.

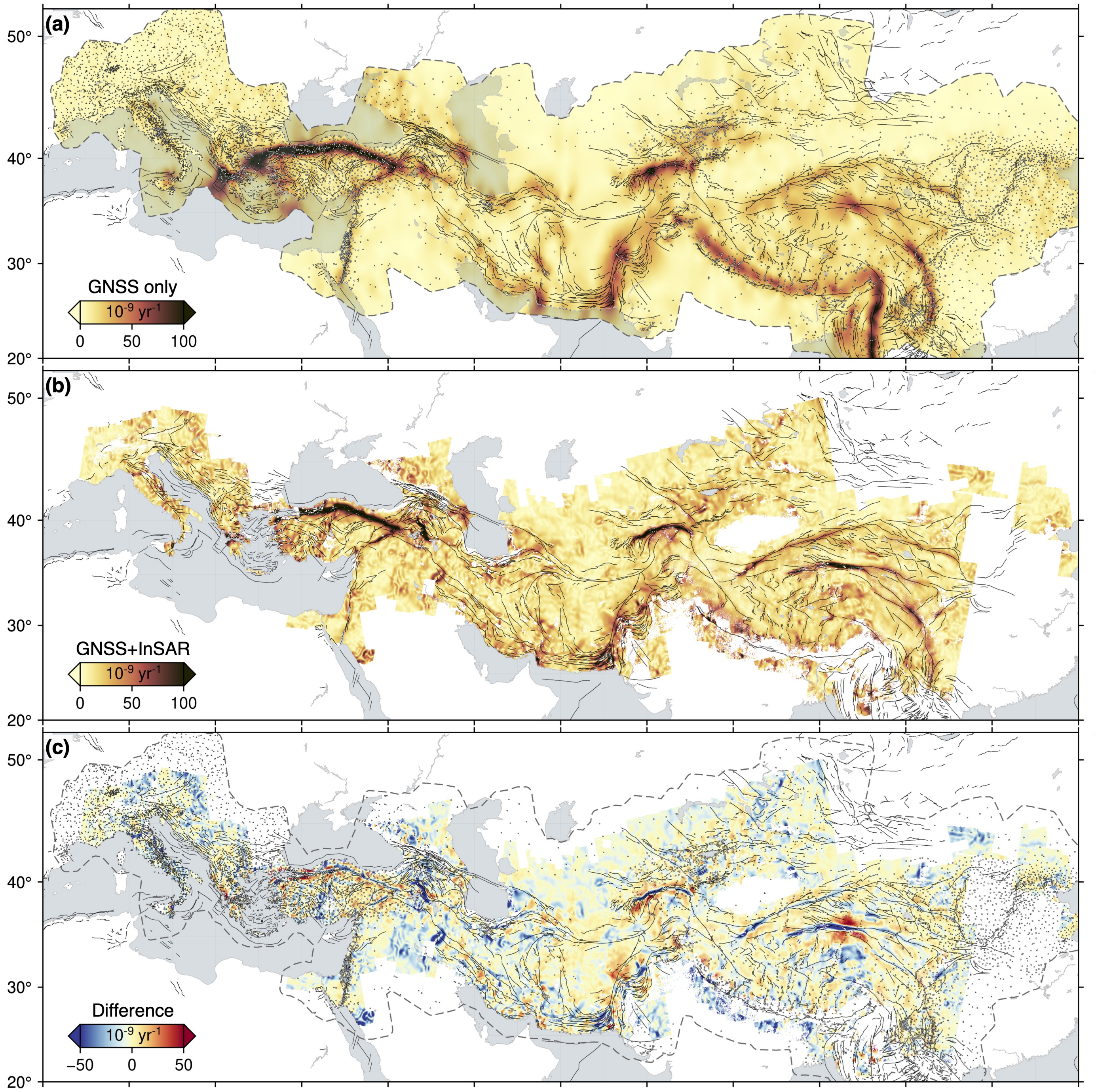

Figure SB29: Comparison of the maximum shear strain rate field derived from (a) GNSS-only versus (b) combined GNSS+InSAR solutions. The difference is shown in panel (c).

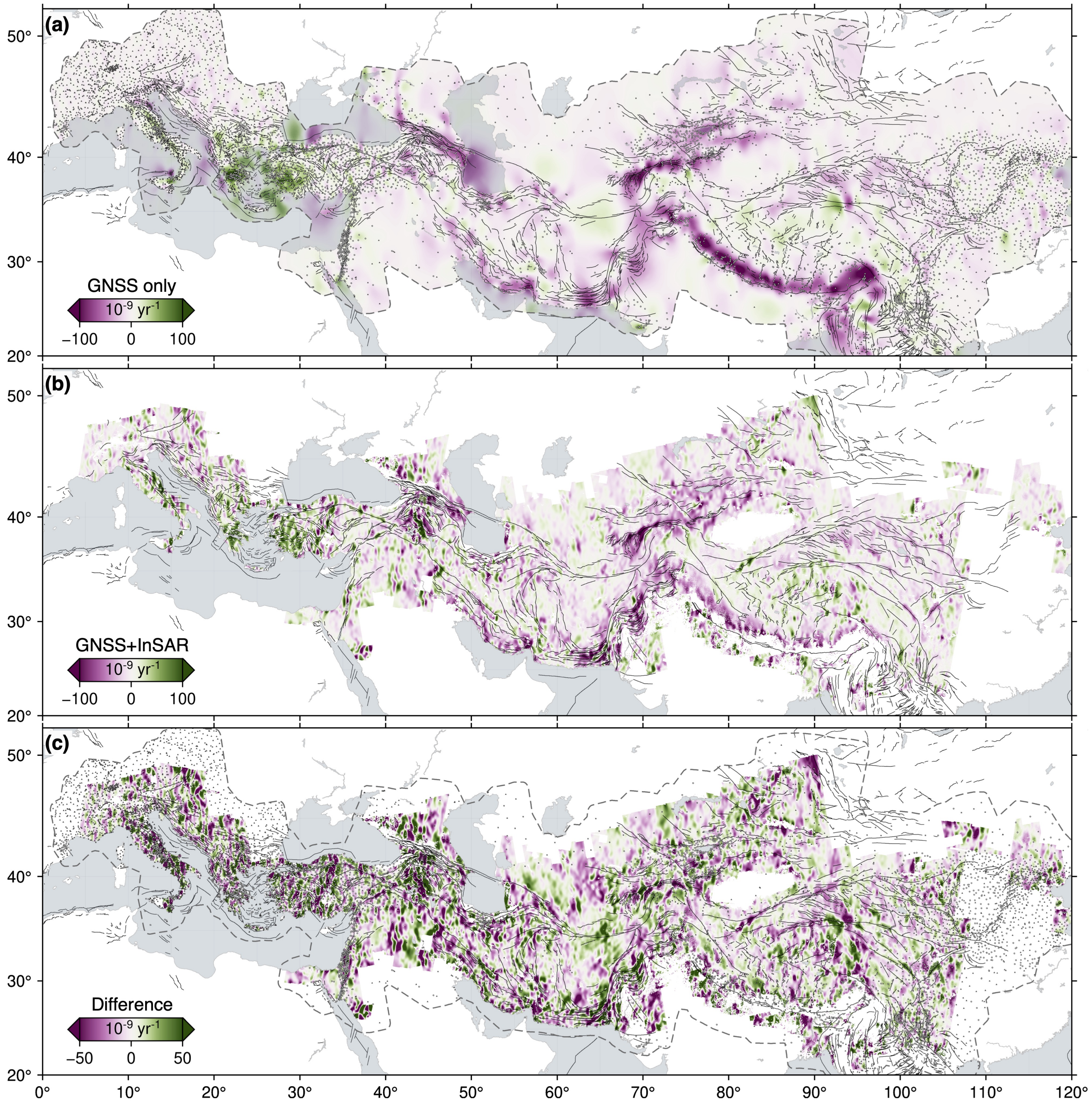

Figure SB30: Comparison of the dilatation rate field derived from (a) GNSS-only versus (b) combined GNSS+InSAR solutions. The difference is shown in panel (c).

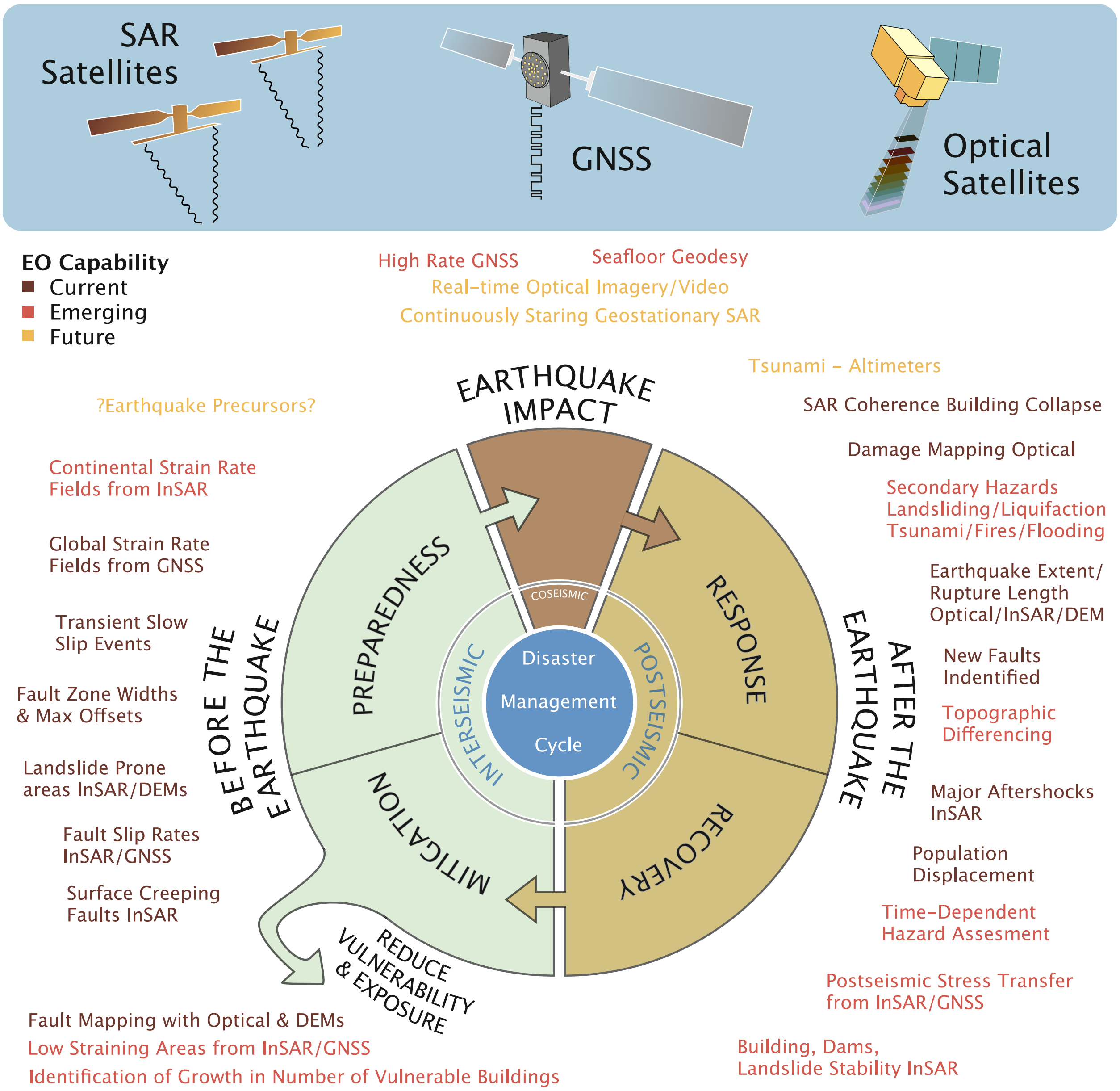

Figure SB31: SAR and GNSS contribute important Earth Observation data constraints (in addition to optical satellites) that provide information to various aspects of the disaster management cycle, such as the emerging field of generating Continental Strain Rate Fields from InSAR combined with GNSS to feed into seismic hazard calculations. Updated from Elliott (2020).

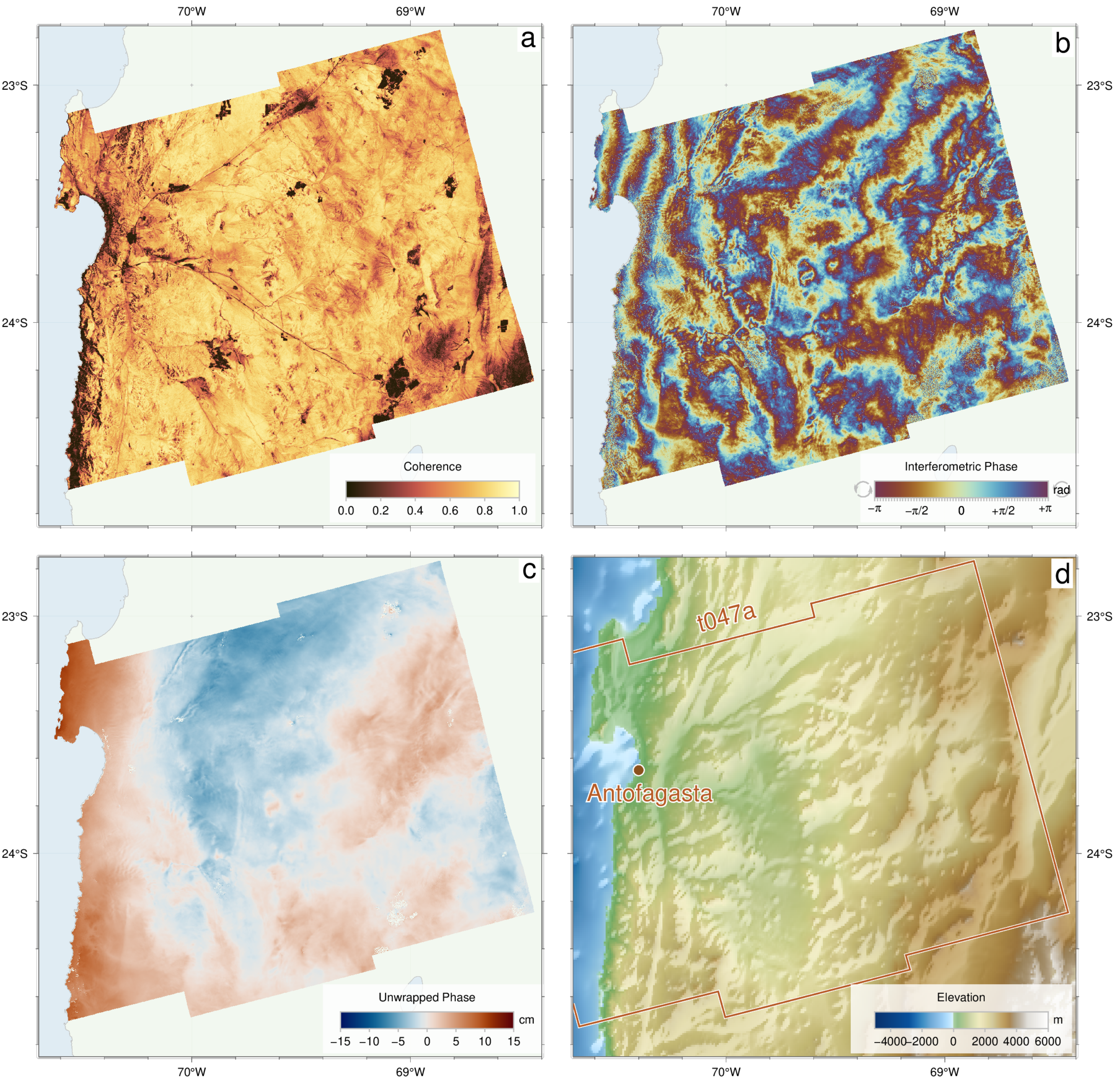

Figure SB32: Outputs from differential InSAR of one of the first Sentinel-1A ascending track 047 acquisitions in this area (20150205) with one of the first Sentinel-1C acquisitions (20250201), spanning an interval of 10 years. The region is between the Chilean Coast at Antofogasta and the Atacama Desert on the Tropic of Capricorn. As one of the driest places on earth, the coherence is maintained very well over 10 years. The unwrapped phase captures both the difference in refractivity of the atmosphere and the ocean tidal loading between the two acquisition dates, but also the cumulative displacement of 10 years of locked elastic loading along this subduction zone. (a) Coherence showing regions that have remained relatively stable (high values) and those were the land surface has changed in the 10-year period (low values, e.g., the coastal fringes in the west, settlements and agricultural patches, linear roads, salars in the east and mines such as in the south-west of the scene. (b) Interferometric phase wrapped $(-\pi, \pi)$ with increasing values indicating range increase) (c) Unwrapped Phase measuring line-of-sight displacement (positive motion, red, indicates motion towards the satellite) (d) Hillshaded Digital Elevation Model of bathymetry and topography at 30 arcseconds resolution from SRTM30_plus (Becker et al., 2009). The outline of interferogram is shown in red.

Appendix C. Supplementary Data

The supplementary material related to this article consists of two csv data tables, one for each of ascending and descending (*AHB.asc.csv & AHB.dsc.csv*) comprising the following information on the SAR frames used in the study:

Frame_ID — consisting of the track number, a unique time and the number of frames on each subswath (typically 13 on each).

TS Date Range — Time interval spanning time series.

#Epochs — Number of epochs (acquisitions) used in inversion for time series.

#Ifgms — Number of interferograms used in inversion for time series.

LICSBAS options run — the flag options and parameter values used in running the LiCSBAS inversion.

Mask Step 15 params — The Masking parameters used in step 15 of LiCSBAS to remove noise and erroneous pixels from the time series.

vstd scaling — the standard deviation uncertainties in the velocity scale by the spatial function given (exponential, linear, spherical) with the values for the nugget, sill and range given).

Earthquake Removal — the choice made whether to solve for an offset or truncate the time series based upon the impact on the frame of coseismic deformation from the list of Earthquakes in Table SA3.

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