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Rapid locking of tectonic magnetic fabrics in weakly deformed mudrocks

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ABSTRACT

The anisotropy of magnetic susceptibility (AMS) has provided important insights into early deformation conditions in compressive settings through characterization of tectonic fabrics in mudrocks that appear otherwise undeformed. The validity of these insights relies on the assumption that tectonic fabrics are rapidly locked shortly after sediment deposition. However, the time lag between deposition and the tectonic overprint has yet to be quantified to verify this assumption. Here we present AMS data from Late Holocene sediments recovered in two cores from Lake Issyk-Kul in the Kyrgyz Tien Shan fold-and-thrust belt. These sediments, in which magnetic fabrics reflect the preferred orientation of phyllosillicates, have typical tectonic magnetic fabrics with varying degrees of tectonic overprint likely controlled by small-scale active faults. The mean orientation of the susceptibility maxima parallels neighbouring active thrust faults and is perpendicular to the geodetically-derived local shortening direction. The age model for one of the studied cores, which is based on 7 calibrated accelerator mass spectrometer radiocarbon dates, indicates that sediments as young as 25 yr record this clear tectonic fabric, which is locked within sediments older than 1180 cal. yr BP. Our demonstration of rapid locking of tectonic fabrics in weakly deformed mudrocks shortly after deposition provides the required validation of their reliability for tectonic studies, and opens new opportunities for providing quantitative insights into the relationship between magnetic ellipsoids, shortening rates, and stress directions in compressive settings.

1. Introduction

Magnetic fabrics have provided insights into early deformation conditions in fold-and-thrust belts (Aubourg et al., 1995; Borradaile and Henry, 1997; Borradaile and Jackson, 2004, 2010; Borradaile and Tarling, 1981; Cifelli et al., 2009; Kissel et al., 1986; Larrasoaña et al., 2004; Mattei et al., 1995, 1997; Parés, 2004; Sagnotti et al., 1998, 1999; Tarling and Hrouda, 1993), foreland basins (Parés et al., 1999; Pueyo Anchuela et al., 2010; Soto et al., 2009; Weaver et al., 2004), and accretionary prisms (Kanamatsu et al., 2001; Kissel et al., 1986; Weaver et al., 2004) through the study of weak tectonic fabrics in mudrocks that typically appear undeformed at outcrop scale. Tectonic analysis of magnetic fabrics of these so-called weakly deformed mudrocks are based on a presumed rapid overprint of deformation on initial sedimentary fabrics shortly after deposition, when mudrocks are still flat-lying, soft and the presence of water allows reorientation of mineral grains according to the prevailing stress field (Benn, 1994; Borradaile and Tarling, 1981; Parés, 2004; Richter et al., 1993). This assumption is particularly important for palaeostress determinations, which require a rapid tectonic overprint so that the magnetic ellipsoid represents the best possible approximation to the incremental strain ellipsoid (Soto et al., 2009). While rapid locking of tectonic fabrics during early burial appears to be well supported (Larrasoaña et al., 2004; Mattei et al., 1995; Parés, 2004; Parés et al., 1999; Sagnotti et al., 1999; Soto et al., 2009), the time lag between sediment deposition and imposition of the tectonic overprint, as well as the time taken for such an overprint to be effectively locked, are yet to be determined. Such quantification has been hampered by difficulties in establishing an absolute chronology for early burial processes in ancient mudrocks, on which most AMS studies have been made. Here we overcome this problem by documenting tectonic fabrics in Late Holocene muds from a tectonically active compressive setting, the Tien Shan fold-and-thrust belt in Kyrgyzstan (Central Asia) (Buslov et al., 2007; Torizin et al., 2009). The time taken by tectonic deformation to overprint the initial sedimentary fabric can be readily constrained by dating the youngest sediments that document a tectonic fabric. On the other hand, the time lag between sediment deposition and tectonic overprinting can be straightforwardly determined by dating the youngest sediments that demonstrably record locked tectonic fabrics.

2. Geological setting

Lake Issyk-Kul occupies the bottom of an E–W oriented depression located between the Kungey and Terzkey mountains in the Kyrgyz

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Tien Shan range (Fig. 1a). This mountain range developed as a result of the far-field effects of the ongoing collision between India and Eurasia, which has led to northwesterly-directed underthrusting of the stable Tarim block beneath the Kazakhstan shield (Buslov et al., 2007; Torizin et al., 2009). The Lake Issyk-Kul basin is filled with Cenozoic continental sediments, and lacustrine sediments have been deposited since the Oligocene (Buslov et al., 2007). The basin is bounded by south- and north-verging basement thrust units that involve Precambrian and Palaeozoic rocks of the Kungey and Terzkey mountains, respectively (Buslov et al., 2007; Torizin et al., 2009). Tectonic activity in the region started in the Oligocene and peaked within the last 3–5 Myr as inferred from apatite fission-track thermochronology and magnetostratigraphic dating of Neogene sediments (Burbank et al., 1999; Buslov et al., 2007). Ongoing deformation has led to several unconformities and displaced geomorphic surfaces within and above Quaternary sediments in the Issyk-Kul area (Bowman et al., 2004). The region is affected by frequent large magnitude (6.5–8.7) earthquakes that occur mainly on the thrust faults that separate the Lake Issyk-Kul basin from the Kungey and Terzkey ranges (Abdrakhmatov and Djanuzakov, 2002; Torizin et al., 2009) (Fig. 1a). Global positioning system (GPS) data indicate a present-day horizontal displacement that ranges between 2 and 10 mm/yr along the northern and southern shores of Lake Issyk-Kul, respectively, relative to stable Asia (Abdrakhmatov et al., 1996; Torizin et al., 2009). This horizontal displacement has occurred along a north–south shortening direction since the Pliocene–early Quaternary (Buslov et al., 2007).

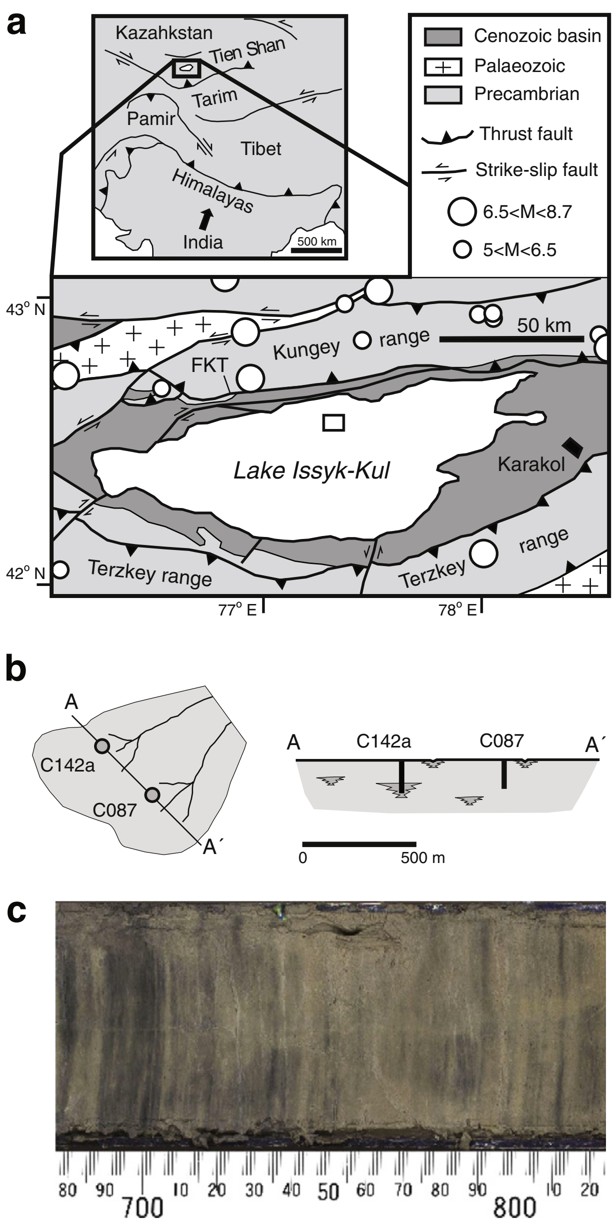

Fig. 1. Geological setting of the studied sediment cores. (a) Geological sketch map of the Lake Issyk-Kul basin and surrounding mountain ranges in the context of the India-Asia collision. FKT denotes the Fore Kungey thrust (Buslov et al., 2007). White circles indicate the location of strong historic earthquakes (Abdrakhmatov and Djanuzakov, 2002; Torizin et al., 2009). The box denotes the location, east of a reverse fault located on the NW shore of the lake, of cores C142a and C087. (b) Sketch of the location of the studied cores within a lobe of the deltaic system that drains the northern shore of Lake Issyk-Kul, in map view and cross-section. (c) Image of sediments that dominate core C087 and most of core C142a. The scale is in millimetres.

3. Materials and methods

Cores C142a (42°34′31.2″ N–77°20′03.0″ E) and C087 (42°34′5.22″ N–77°20′10.44″ E) were collected using a gravity corer at water depths of 150 and 312 m, respectively (Fig. 1a). These azimuthally unoriented cores were recovered from the upper slope of the central northern margin of Lake Issyk-Kul, just east of a reverse fault located on the NW shore of the lake. Core C142a comprises a 150-cm-thick sequence of Late Holocene clays, silts and sandy silts that accumulated on a distal lobe of the deltaic system that drains into the lake along a NE–SW direction (Fig. 1b) (see de Batist et al., 2002). The uppermost 18 cm of the core is composed of dark, organic-rich and massive silty clays. Between depths of 18 and 94 cm, the sediments consist of alternating light and dark grey clays. The lowermost 56 cm of the core consists of alternating light and dark silts and sandy silts that provide evidence for a distributary channel that supplied coarser-grained sediments to the deltaic system. Core C087 comprises 132 cm of clays and silty clays that are equivalent to those in the uppermost 94 cm of core C142a (Fig. 1c). The lack of coarser-grained sediments at the base of core C087 might be explained by the discontinuous nature of deltaic tributary systems (Fig. 1b).

Back-to-back palaeomagnetic samples were obtained by pushing 63 and 52 plastic boxes (2×2×2 cm) into the working half of cores C142a and C087, respectively. AMS measurements were conducted using an AGICO KLY-2 magnetic susceptibility meter at the Paleomagnetic Laboratory of the ICT Jaume Almera (UB-CSIC), Barcelona, Spain, using a field of 0.1 mT and a frequency of 470 Hz. The AMS is a second-rank tensor that can be graphically displayed as a three-axis ($\kappa_{max}>\kappa_{int}>\kappa_{min}$) ellipsoid with a given orientation, shape and degree of anisotropy (Tarling and Hrouda, 1993). The shape and degree of anisotropy of the magnetic ellipsoid for each sample can be described using the shape (T) and corrected anisotropy degree (P^j) parameters of Jelínek (1981), respectively. T ranges between 1 and −1 for pure oblate and prolate ellipsoids, respectively, and is about 0 for triaxial ellipsoids (Jelínek, 1981). P^j gives an idea of the strength of the ellipsoid (Jelínek, 1981). Its value depends on the degree of anisotropy of the rock-forming minerals, and ranges between 1 and 1.15 for weakly deformed mudrocks whose magnetic susceptibility is dominated by paramagnetic minerals (Parés, 2004). Relevant AMS parameters for each sample are summarised in the supplementary data as a function of depth and age. The mean orientation of the magnetic ellipsoid for all samples and the associated confidence ellipses were calculated following Jelínek (1981).

Palaeomagnetic analyses were conducted at UB-CSIC, and involved progressive alternating field (AF) demagnetization and subsequent measurement of the natural remanent magnetization (NRM) for all samples. The NRM was measured using a 2-G Enterprises three-axis superconducting rock magnetometer, with AF demagnetization up to a maximum field of 100 mT at steps ranging between 5 and 30 mT. Stable palaeomagnetic directions were identified from orthogonal demagnetization plots (Zijderveld, 1967) and were calculated by