Geological and Seismological Field Observations in the Epicentral Region of the 27 September 2003 M_w 7.2 Gorny Altay Earthquake (Russia)

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Abstract The $M_{\rm w}$ 7.2 Altay (Chuya) earthquake of 27 September 2003 that occurred in Gorny Altay (southern Siberia, Russia) is the first event of this magnitude to occur in the region in historical time. The 60 km long surface ruptures of the rightlateral event follow a preexisting but previously unmapped northwest-southeast fault trace along the northern slope of the North Chuya range. Additional secondary coseismic ruptures were also observed on adjacent faults. The earthquake triggered landslides, rock falls, and liquefaction, as well as destruction of houses and other construction in the Kurai and Chuya basins. The 2003 event induced several thousand aftershocks during the years after the mainshock that formed a cloud of epicenters aligned with the main coseismic rupture trace. The local earthquake tomography of the event source shows the correspondence between the fault trace at the surface and a long narrow low-velocity zone, penetrating vertically into the crust down to a depth of 15–17 km. If the three-dimensional (3D) geometry of the aftershock cloud approximates the ruptured fault plane at first order, then the fault is mainly vertical, or slightly southwest dipping, confirming the right-lateral-reverse kinematics of the fault, compatible with northeast-southwest shortening accommodated by a combination of right-lateral and thrust faults. Together with the average surface displacement observed in the field (Δu 1–2 m), the surface of the fault plane determined by the aftershocks distribution (80 × 17 km) gives a magnitude $M_{\rm w} \sim 7.2$, in good agreement with the Harvard determination. Natural exposures produced by the 2003 surface faulting, together with previous paleoseismic observations across the primary and secondary earthquake-induced features, have revealed the occurrence of several strong events (magnitudes about 7–8) during the last 5000 yr. The characteristics of the Altay event suggest that the Gorny Altay region, similar to Mongolia and Gobi, is characterized by large infrequent M 7–8 earthquakes along faults moving at rates of a few mm/yr or less.

Introduction

The M_w 7.2 earthquake that occurred in southern Siberia, in the Russian part of Gorny Altay near the border of Kazakhstan, China, and Mongolia, on 27 September 2003, is the strongest event historically reported in this region (Rogozhin *et al.* 2003; Nissen *et al.*, 2007; Fig. 1). This almost pure right-lateral strike-slip event occurred along a previously unrecognized fault trace and was associated with spectacular ground failures including primary surface ruptures, very large open fissures, and landslides. It was followed by three strong aftershocks: an $M_{\rm s}$ 6.3, 25 min after the mainshock, an $M_{\rm w}$ 6.4 on 27 September 2003, and an $M_{\rm w}$ 6.6 on 1 October 2003 (Fig. 2, Table 1). The poor regional coverage by seismological stations resulted in low-quality locations for these events by both regional and worldwide seismological centers (Fig. 2).

For a decade, the Schmidt Institute of Physics of the Earth (Russian Academy of Sciences) has performed geological studies in the region (Rogozhin *et al.*, 1998, 2003, 2004, 2007). The paleoseismic investigations on secondary seismic features such as liquefaction and landslides completed before the earthquake, in 1996 and 1998, suggested the occurrence of large seismic events, although the faults

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Figure 1. Regional seismotectonic map of the Altay in Siberia and western Mongolia (modified from Schlupp [1996] and Rogozhin *et al.* [2003]). Main active faults and historical strong earthquakes are presented. The inset shows the location of this map within Eurasia. The black rectangle represents the location of Figures 2 and 3. (1: strike-slip faults, 2: reverse faults, 3: normal faults, 4: surface rupture traces of historical large earthquakes, 5: location of the 2003 Gorny Altay earthquake, 6: region above 2000 m).

directly responsible for these shaking earthquake-induced features had not been identified. Based on these results, the Russian Altay was classified as a region of high seismic risk and is shown on the Russian Federation seismic zonation as a zone with expected earthquake intensity up to IX on the European intensity scale MSK64 (Strakhov and Ulomov, 1999). Even before 1997, this region was referred to as a zone of expected intensity up to VIII (Strakhov and Ulomov, 1999). The major regional strike-slip fault, the Kurai fault (KF), which runs north of the Chuya and Kurai basins (Fig. 3), is responsible for large (≥ 1 km) dextral river offsets, although it has been described as left lateral from field observations of fault outcrops (Delvaux et al., 1995; Thomas et al., 2002). It is the clearest tectonic feature in the landscape, but shows almost no seismic activity. The 2003 event occurred south of the basins on the northern slope of the Chuya range where no fault had been identified before the earthquake (Fig. 3).

Several expeditions visited the epicentral zone of the earthquake for seismological, geological, and paleoseismological studies in the autumn of 2003 (Rogozhin et al., 2003) and during the summers of 2004 (Areviev et al., 2004; Goldin et al., 2004; Rogozhin et al., 2004; Lunina et al., 2008) and 2005 (Rogozhin et al., 2007). The main seismic rupture traces were recognized and mapped, and we found that the dextral surface rupture extends for as much as 60 km. Strong ground motions during the mainshock generated many secondary gravitational and vibrational breaks (Fig. 3). We installed up to 12 mobile seismic stations in the epicentral zone during the summers of 2004 and 2005. The processing of the seismological data and the good quality of the digital records allow us to use a double-difference tomography method to study the velocity structure and to obtain precise hypocentral locations, allowing us to map aftershocks and show several cross sections along and across the source area. In this study, we present and compare our



Figure 2. Locations of mainshock and eight major aftershocks, as given by various seismological centers. Different colors correspond to different events, and different symbols refer to different seismological centers. Drawn on top are Harvard centroid moment tensor (CMT) solutions for these events (Table 1). Black symbols correspond to the 27 September 2003 M_s 6.3 event for which no mechanism is determined. Coseismic rupture trace is shown on top of shaded relief map.

field geological and seismological observations for a better understanding of the source of the 2003 earthquake.

Tectonic and Geological Setting

The source area of the earthquake is situated in the reactivated ancient Caledonian fold system of Gorny Altay (Tapponnier and Molnar, 1979; Cunningham *et al.*, 1996; Fig. 1). In the epicentral area, crystalline and low-grade metamorphic Paleozoic rocks, mainly Silurian schist, phyllite, and marbles, are deformed in narrow linear folds that make up large anticlinoria and synclinoria (Delvaux *et al.*, 1995). Tertiary deformation began in the Pliocene long after the onset of the India–Asia collision (Tapponnier and Molnar, 1979; Cun-

Date (dd/mm/yyyy)	Time (hr:min:sec)	Latitude (°)	Longitude (°)	$M_{\rm s}$	$M_{\rm w}$	$M_0\times 10^{16}~({\rm N~m})$
27/09/2003	11:33:36	50.02	87.86	7.5	7.2	9380
27/09/2003*	11:58:07	49.98	88.09	6.3	_	_
27/09/2003	18:52:53	50.09	87.75	6.6	6.4	450
01/10/2003	01:03:30	50.24	87.59	7.1	6.6	1130
09/10/2003	16:06:03	49.75	88.05	4.5	5.0	4.13
13/10/2003	05:26:42	50.25	87.75	5.0	5.1	6
17/10/2003	05:30:26	50.27	87.94	4.7	5.1	6.05
23/10/2003	00:25:48	49.64	88.16	4.8	5.1	4.84
11/11/2003	22:42:36	50.47	86.97	4.4	5.1	5.26
17/11/2003	01:35:52	50.24	87.53	4.7	5.2	7.53

 Table 1

 Main Events Characteristics of the 2003 Sequence (Source: Harvard)

*No Harvard CMT solution for this event. Data are from the ISC.



Figure 3. Seismotectonic map of epicentral area. Main surface rupture (bold red lines) of the 2003 Gorny Altay earthquake follows the northern edge of the Chuya range along the North Chuya fault (NCF). The Chuya and Kurai basins filled with Tertiary sediments are limited north and south by growing ranges: the Kurai and Chuya mountains, respectively. The Kurai fault (KF), the major right-lateral strike-slip fault in the area, was not activated during the 2003 earthquake. We distinguish three rupture sections of the 2003 surface rupture (I, II, and III): the northwestern section along Kurai basin, the central section where the fault crosses the saddle between the Sukkor block and the Chuya range, and the southeastern section where the rupture splays into several branches (see the Discussion section of the text). Northwest of the Kurai basin, ground breaks (thin red lines) may be secondary shaking-induced features, although they are associated with preexisting fault scarps (Rogozhin *et al.*, 1998). Focal mechanism and magnitude from Harvard are also shown. SKF stands for the South Kurai fault.

ningham *et al.*, 1996) when the region, peneplaned during the late Paleozoic and Mesozoic, began to rise due to combined thrusting and strike-slip faulting (Delvaux *et al.*, 1995; Cunningham *et al.*, 1996).

The Tertiary deformation of the region resulted in the formation of ranges separated by deep intermontane basins. The 2003 epicentral area is bordered by the Aygulak–Kurai range to the north and the Chuya–Saralyugem range to the south (Fig. 3). These northwest–southeast striking ranges are separated by the Kurai and Chuya basins, northwest and southeast, respectively, of the moderately elevated Sukkor (Chagan–Uzun) block. Paleozoic (Caledonian) rocks outcrop in the ranges, partly covered by Quaternary moraines.

In the basins, soft Neogene deposits, Quaternary moraines, lacustrine, and alluvial deposits are accumulated (Baker *et al.*, 1993). The thickness of the Tertiary and Quaternary deposits reaches 500–600 m and 1000 m in the Kurai and Chuya basins, respectively (Thomas *et al.*, 2002).

During the Pleistocene, large parts of the Altay region were covered by glaciers and ice caps that reached a thickness of up to 2 km (Rudoy, 2002). During the last glaciation, the Kurai and Chuya basins were surrounded by ice that dammed the Chuya river outlet to the northwest, resulting in an impounded lake. The highest lake shoreline is situated at about 2100 m above sea level, about 400 m above the basin bottom and probably postdating the last glacial maximum (LGM, ~20 ka). One of the largest worldwide cataclysmic floods resulted from the breakage of the ice dam during deglacation about 16 ka, as determined by ¹⁰Be surface exposure dating (Baker *et al.*, 1993; Rudoy, 2002; Herget, 2005; Reuther *et al.*, 2006). Still, after the lake disappeared, some of the largest glaciers continued to flow into the basin as recorded by frontal moraines in the central part of the Chuya basin, probably due to an increase of moisture during the Pleistocene–Holocene transition (Rudoy, 2002; Fig. 3). The landscape of the epicentral area has thus been intensively impacted by glacial and periglacial erosion, and also by the late glacial lacustrine environment. Therefore, the imprint of seismic activity in the geomorphology of the piedmonts probably only spans the last 15 ka; older traces of deformation were certainly smoothed out by glacial erosion.

Nevertheless, several active faults are recognized and mapped that mainly follow the basin margins on the south and north (Fig. 3; Rogozhin *et al.*, 2003). The most prominent are the Kurai fault (KF), the South Kurai fault (SKF), and the North Chuya fault (NCF) that trend mostly westnorthwest and northwest (Fig. 3), although the NCF was not recognized before the 2003 earthquake. Other minor faults affect the northwestern margin of the Kurai basin (Fig. 3).

The SKF zone extends along the southwestern margin of the Kurai range and follows the northern limit of the Kurai and Chuya basins. North of Kurai city, it is associated with a chain of small-scale northwest-trending anticlines bounded by active reverse faults, suggesting ongoing northeastsouthwest-directed shortening at the northern margin of the Kurai basin. To the southeast, folding and thrusting affects the piedmont of the Kurai range northwest of Kosh-Agach city. About 10 km northeast of the SKF zone and aligned with the highest part of the Kurai range crest line, the KF is a clear strike-slip fault, marked by a series of right-lateral stream offsets ranging from 4 to 10 km (Fig. 3). The NCF is situated southwest of the basins on the northern slope of the Chuya range, crosscutting the ancient glacial valleys of the North and South Chuya ranges. It is poorly expressed in the landscape and is recognized as small discontinuous scarps and saddles along the slopes. The 2003 Altay earthquake occurred along this fault.

The well-known active faults on the territory of western Mongolia and northwestern China extend towards the Russian Altay (Fig. 1). In these regions, several large seismic events with magnitudes $M \ge 7.0$ occurred in the twentieth century (1905 Bolnay, 1905 Tsetserleg, 1931 Fu Yun, 1957 Gobi-Altay; Fig. 1) as well as earlier (1761 Ar-Hotol, Baljinnyam *et al.*, 1993; Fig. 1). In the existing stress field (Dricker *et al.*, 2002), the faults of northwest–southeast or north-northwest–south-southeast strike are dextral with a small normal component, and the faults with near east–west orientation are sinistral or reverse (Schlupp, 1996). The kinematics of coseismic movements along the primary fault system of the 2003 Altay earthquake is compatible with the regional tectonic regime.

Field Geological Observations

The preliminary geological field survey of the source area was carried out a few days after the mainshock in the late autumn of 2003 (Rogozhin et al., 2003). Further investigations of the entire epicentral zone took place during the summers of 2004 (Rogozhin et al., 2004) and 2005 (Rogozhin et al., 2007). We observed primary coseismic rupture traces, as well as secondary gravitational and vibration dislocations in mountains and foothill parts of the epicentral area (Fig. 4). Secondary features include landslides, rock falls, debris, slope instabilities in the steep parts, liquefaction, mud volcanoes, and subsidence, which are all induced by shaking. Most of these features are characteristic of large M > 7 strike-slip events as described along other recent major ruptures worldwide such as the M_w 7.9 14 November 2001 Kokoxili event in Tibet (Lin et al., 2004; Klinger et al., 2005) and the M_w 7.9 3 November 2002 Denali event in Alaska (Eberhart-Phillips et al., 2003; Harp et al., 2003).

During the fieldwork, some paleoseismic information was also obtained in the epicentral area of the 2003 earthquake. At several places, the rupture breaks provided natural exposures across colluvial deposits along preexisting scarps. Samples of organic material suitable for radiocarbon dating allowed us to estimate the time of the ancient earthquake occurrence at the same location (Rogozhin *et al.*, 2003, 2007). This dataset complements the paleorupture dating completed earlier in the same area (Rogozhin *et al.*, 1998).

The Main Coseismic Rupture

The seismic source of the 27 September 2003 earthquake forms an almost linear system of primary ground ruptures (Fig. 4) that reach a total length of about 60 km (Fig. 3). The rupture orientations are generally northwest (sometimes west-northwest). They follow the preexisting trace of the NCF along the northeastern slope of the North Chuya and South Chuya ranges and crosscut all of the relief. The strike of the fault rupture correlates with the orientation of the long axis of the aftershock cloud, N132° E (Arefiev *et al.*, 2004). The densest concentration of epicenters lies exactly in the zone of seismic fault ruptures. Maximum dextral offset is about 1.5–2.0 m. In some places, the vertical component of displacement reached 0.7 m (Figs. 4c and 5).

Following the morphological characteristics of the rupture traces, the main rupture is divided into three sections: the northwestern, central, and southeastern sections that are labeled I, II and III, respectively, in Fig. 3.

Northwestern Section

The northwestern section of the rupture zone was investigated along the watershed of the Akturu and Mazhoy rivers (Fig. 3). The northwest termination of the rupture is observed in the Mazhoy valley, where the fluvio-glacial deposits show small open fissures and ridges with amplitudes of offsets



Figure 4. Field views of ground rupture of the M_w 7.2 Gorny Altay earthquake. Locations of the photographs are shown in Figure 3. (a) View to the northwest of mole tracks and open fissures south of the Taldura valley (photo is courtesy of A. R. Geodakov). In the background, arrows point to the two branches of the rupture. (b) Rupture trace divided into two parallel branches with total coseismic offset of about 50 cm, north of the Taldura valley. Black arrows indicate right-lateral movement. White arrows indicate edges of the road that are offset by the two main branches of the rupture. (c) View to the northwest along the northern branch of the southeastern section. The 2003 surface rupture is associated with meter size vertical displacement (steep free-face scarp indicated by a white arrow), rejuvenating base of preexisting scarp (smooth slope indicated by black arrow). (d) View to the south of the largest coseismic landslide in the epicentral area of the mainshock. Coseismic ruptures are visible just above the cliff upslope of the landslide and at its base (white arrows). [Photos (b), (c), and (d) are courtesy of A. N. Ovsyuchenko.]



Figure 5. Amplitude of right-lateral and vertical offset as measured along three sections of the 60 km long rupture. Measurements are projected along the strike of the rupture. Positions of main watersheds are indicated as blue lines.

ranging from 0.1 to 0.3 m, both vertically and horizon-tally (Fig. 5).

Central Section

Farther to the southeast, the central segment of the rupture (Fig. 3) is developed in upper Pleistocene glacial deposits, which overlie the Middle Devonian shale. Fractures appear in rocks as en echelon sequences of fissures with slickensides on the walls, both ancient and fresh. Orientations for both systems of slickensides are similar. Dextral offsets are about 2 m and the vertical offset component reaches 0.5 m at one place (Fig. 5).

The most impressive seismic ruptures were observed in the 4 km wide saddle on the watershed divide of the Taldura and Kuskunnur rivers (Lunina et al., 2008). There, the zone of coseismic faulting is developed in moraine deposits and is comprised of en echelon fractures that trend nearly eastwest. The fissures have lengths up to 300 m, widths of 10 m, and depths up to 30 m, and they alternate with linear pressure ridges. The total width of the rupture zone is about 200-250 m. The maximum dextral offset measured is 1.5 m (Fig. 5; Rogozhin et al., 2003, 2007). Pressure ridges have heights up to 2 m and lengths of 50 m. The ridges are accompanied by thrusting, causing soil layer doubling. These pressure ridges are similar to those observed along the $M_{\rm w}$ 7.9 Kokoxili earthquake that occurred in November 2001 when the rupture broke frozen ground atop a permafrost layer (e.g., Klinger et al., 2005; Xu et al., 2006).

Large pull-aparts are observed on the slopes of the saddle, with widths reaching 50–500 m. They are bordered by cumulative scarps with heights of 2.5 m that were reactivated during the 2003 earthquake. On the walls of the fissures, Rogozhin *et al.* (2003, 2007) observed slickensides and zones of crushed rocks and sediments. They also observed multiple layers of ancient soils inside of the fissures, which are preserved probably because the refreshed steep

scarps collapsed onto the soil after previous large earthquakes. Rogozhin *et al.* (2003) interpret the sequence of soil layers separated by gravel horizons to represent a sequence of past events at about 3000, 1500, and 1000 yr ago.

In the Kuskunnur river valley, the fault zone formed an en echelon system of open fissures between which orthogonal ridges are formed (Rogozhin *et al.*, 2007). Dextral offset measured there was 1.5 m (Fig 5). The trunk of a tree had split, and one part shifted relatively to the other.

A pressure ridge cut by the 2003 rupture offers a natural exposure to a depth of 2.5 m (Rogozhin *et al.*, 2007). Two paleosols are recognized on the exposure, with the deeper one showing more deformation. The deeper layer is almost duplicated in an overturned fold, which thrusts onto the downthrown side block. The shallower buried paleosol is less deformed but is overthrusted. Finally, these deformed horizons were abraded and covered by the present soil horizon and faulted during the 2003 event. Rogozhin *et al.* (2007) interpret these observations, together with radiocarbon dating of some of the soil layers, to imply the occurrence of at least two events in the last 1000 yr and maybe three events in the last 5500 yr.

Southeastern Section

Near the southeastern end of the main fault, the rupture bifurcates into two north-northwest-striking subparallel branches that are up to 5 km apart (Fig. 3). We observed dextral offsets on both branches. Between them, a pull-apart-like structure has formed with 0.2–0.3 m of subsidence (Fig. 5). East–west-striking pressure ridges, 0.3–0.7 m high, developed in this local depression. It is important to note that the surface between the two branches is extremely broken due to the ground shaking during the earthquake. The slopes were covered by numerous landslides and rock falls (e.g., Fig. 4d). Even large boulders were moved or thrown up during the mainshock. Some parts of the southeastern section

of the fault rupture are characterized by en echelon series of open fissures. Some cracks are up to 1-3 m wide with a length of about 150 m.

The southeastern extremity of the rupture is expressed by a separate branch of fissures in the southern part of the Chuya basin. Normal faulting is typical of this section, especially within the Cambrian and Ordovician shale outcrops. The northern side of the fault subsided by 0.6–0.7 m (Fig. 5). The strike-slip component is shown by an en echelon system of open fissures with orthogonally oriented pressure ridges in between glacial and alluvial deposits. The dextral offset reaches 1.2 m (Fig. 5).

Farther to the southeast, the fault sharply changes strike to become east–west. In the Irbitsu riverbed, small en echelon pressure ridges are bounded by small reverse faults, with no extensional features. The dextral offset there is 0.6-0.7 m (Fig. 5).

Other Faults Activated during the 2003 Earthquake

Paleoseismic studies in the Kurai and Chuya basins and surrounding ranges prior to the earthquake (Rogozhin *et al.*, 1998) revealed that the anticline ridges north of Kurai in the northern parts of both basins are active structures (SKF, Fig. 3). These ridges are bordered north and south by well-developed scarps. Trenches excavated across both scarps exposed faults near the bases of the ridges. During the 2003 earthquake, one of these scarps was reactivated. In the upper part of the scarp, a new 20–30 cm high steep scarplet formed. Open fissures (3–5 cm) appeared at its base. The reactivated scarp strikes northwest and was observed along a distance of about 5 km (Rogozhin *et al.*, 2007).

A similar system of east–west ruptures appeared as a result of the 2003 earthquake along the southern slope of the Kurai range, west of Kurai (thin red lines, Fig. 3). The new coseismic scarps are about 20–40 cm high. The chain of reactivated scarps continues to the west into the canyon of the Chuya river near the Mazhoy river mouth, but does not join with the main fault zone on the northern slope of the North Chuya range. Whether these isolated ground ruptures are primary or secondary rupture traces that result from the mainshock or one of the largest aftershocks remains uncertain. These ground ruptures, however, may be triggered slip from the mainshock on nearby active faults.

Secondary Coseismic Features

Many secondary coseismic features appeared in the epicentral zone during or shortly after the mainshock (Fig. 3). Gravitational and vibration ruptures, like landslides, rock falls, and cracks were widely distributed in the Chuya and Kurai basins and in the surrounding ranges and hills. All of these ruptures are situated inside of a zone about 70 km long and 15 km wide, oriented northwest–southeast (Fig. 3). The area of the secondary ruptures generally correlates with the location of the aftershock cloud. The density of the secondary features in the epicentral area decreases with distance from the primary coseismic ruptures.

The most impressive secondary feature is the great landslide that occurred on the right bank of the Taldura river near the Beltir settlement (Figs. 3 and 4d). This landslide formed northeast of the primary surface rupture, near the transition between the central and southeastern sections of the rupture. This transition is marked by both a change in strike from northwest–southeast to north-northwest–south-southeast and the separation of the rupture into two branches. The body of the landslide moved down from its source by about 150 m and by about 100 m horizontally towards the Taldura valley. The volume of the slide mass is about 30 million cubic meters. Two huge ancient landslides are situated near the gravitational structures of the 2003 earthquake. Several landslides of smaller size also occurred in 2003 on slopes near the epicentral area.

A wide territory was covered by rock falls from steep and vertical slopes in the mountains. Some boulders were as large as a single story house. There were many manifestations of slope instability such as sliding of loose material and turf.

A striking manifestation of shaking at the surface is the formation of liquefaction features in lowland swamps. Oversaturated, loose, thin sediments (sand, clayey sand, silty sand) are widespread in the geological sequence of these landscapes. Many cracks, 0.5-3.0 m wide and up to 50 m long, are associated with liquefied sand and gravelly sand, as well as mud volcanoes. Corresponding subsidence of the surface also occurred in zones with intensity higher than VII. Many indications of liquefaction were observed in the valleys of the Chuya and Chagan-Uzun rivers, and small riverbeds were flooded in the area near the fault rupture. Our trenching study into young depositional sequences in several places allows the identification of past liquefaction features. In the trenches, the liquefaction under the modern turf is evident as a thin white layer. At some places, this sand horizon covers a buried paleosol. Whether these past liquefaction traces are due to earlier earthquakes on NCF can only be determined by cross correlation with faulting evidence on primary ruptures and needs further investigation (e.g., Rogozhin et al., 1998, 2003).

We observe evidence of large boulders jumping and rotating at the time of the mainshock near the epicentral zone, implying ground acceleration exceeding 1g.

The secondary ruptures are irregularly distributed in the epicentral area and are concentrated in narrow linear zones along steep slopes for gravitational features and near rivers, swamps, and lakes with shallow underground water table for liquefaction features.

Field Seismological Observations

We installed a temporary seismological network (Fig. 6) of 12 stations during the summer of 2004, which was partly reinstalled during the summer of 2005. Here we present only



Figure 6. (a) Map of well-located aftershocks (rms < 0.2). Black triangles are seismic stations. Stars are positions of main aftershocks from the International Seismological Centre (ISC) (see Fig. 2). (b) Frequency-magnitude plot. (c) Histograms of main parameters of the earthquake catalog: hypocentral depths, rms on solutions, and errors on depth and on horizontal coordinates of event location.

the seismological data of 2004. Indeed, in 2005, a smaller number of stations had been installed while the aftershock activity, although still very high, was less important, and therefore, no tomography and relocations of equivalent quality could be processed and compared to the data of 2004. From June to September 2004, our temporary network recorded about 1700 aftershocks. Because the southwestern part of the epicentral zone is practically out of reach (summits of the Chuya range), the geometry of the network is not ideal; most of the stations are concentrated northeast of the earthquake rupture.

We used four homemade autonomous 24-bit recorders and two types of short-period three-component sensors, passive (SM-3) and with electronic feedback (KMV). The eight other stations were radio-telemetered stations, four with three-component sensors (KMV) and four with onecomponent sensors (SM-3). This network operated continuously. In the field, we interpreted the data from the telemetric system day by day. Data from the autonomous stations were collected every 5–7 days. We stored the data at the field center of data collection (Kurai) and preprocessed them for interpretation. At the beginning of the experiment, we had no reliable information about the velocity model in the source zone; thus, we made our preliminary interpretation using a half-space model with a 5.5 km/sec *P*-wave velocity.

Following the fieldwork, all data collected were reinterpreted using the program SEIPICK developed in the laboratory of strong earthquakes (Institute of Physics of the Earth [IPE], Russian Academy of Sciences [RAS]). The waveform parameters, such as P- and S-wave arrival times, maximum P and S amplitudes, duration of records, and polarity of *P*-wave first motion, were measured from the records. Using the VELEST program (Kissling et al., 1994), we constructed a one-dimensional (1D) velocity model of the source zone after numerous runs (Table 2). Our data allow us to estimate the velocity down to a depth of 18 km. At greater depth, we used a model based on results from deep seismic sounding (E. E. Zolotov, oral communication, 2004). Using this velocity model and station corrections, we recalculated all earthquake locations with HYPO71 (Lee and Lahr, 1975). A total number of 1694 earthquakes are located. We used different trial starting depths to improve the estimates. In principle, VELEST allows location estimates, but we prefer to use HYPO71 because its output is more convenient for the next step of the study, that is, the construction of focal mechanisms. In addition, for the final location we will use tomoDD (Local Earthquake Tomography [LET]) results (see the next section).

Record durations were measured for magnitude evaluation. However, this method does not provide precise results. As our instruments were well calibrated, we decided to use, for the final catalog, the local magnitude obtained through the amplitudes of P and S waves:

Velocity Model Used in Inversion					
Layer	V _p (km/sec)	H _{top} (km)			
1	4.50	0.00			
2	5.67	2.00			
3	5.83	4.00			
4	5.86	8.00			
5	6.01	10.00			
6	6.25	13.00			
7	6.56	18.00			
8	6.70	23.00			
9	6.80	32.00			
10	6.90	37.00			
11	7.10	42.00			
12	8.20	51.00			

 Table
 2

 Velocity Model Used in Inversion

$$M_1 = \log(A) + b * \log(R) - C,$$
 (1)

where b = 0.8, C = 0.5, A is the sum of $A_{P \max}$ and $A_{S \max}$ in μ m, and R is the epicentral distance in kilometers. This value is well correlated with the best estimate obtained using the duration.

The map of best located earthquakes (rms ≤ 0.2) is plotted in Figure 6 (N = 1578) together with histograms showing the main parameters of the catalog (Fig. 6c). The frequency-magnitude plot is also shown in Figure 6b. As seen, the set of data is complete for $M_1 \geq 1.25$. This low value confirms the high sensitivity of the instruments and of the quality of the sites where they were installed. For this representative part of the catalog (N = 1228), the slope calculated using the maximum likelihood method is equal to -0.947 ± 0.014 . The least-squares method gives a value of -0.977 ± 0.015 .

Because of weather conditions, the field campaign took place 8 months after the mainshock. Magnitudes of aftershocks versus time from the mainshock are drawn using data from the National Seismological Network (Fig. 7a) and from our temporary network (Fig. 7b). The level of seismic activity decreases very slowly with time (Fig. 7b) and allows late field seismological observations. There was still a high activity 10 months after the mainshock and, moreover, locally collected data are much more precise than permanent network data.

Several M 4–5 earthquakes occurred during our field experiment in the summer of 2004 (Fig. 7), which gave us an opportunity to test the quality of locations given by various regional, national, or international seismological networks. Figure 8 shows the difference between our locations, which we assume are very close to the real position of the corresponding events, and the locations given by the NEIC, OBN (national network), and NVS (regional network). The average location differences for these networks are 20.9, 16.2, and 17.2 km, respectively. These values give us a very rough estimation of the location errors for the catalogs.

We show in Figure 9 69 focal mechanisms built following Rivera and Cisternas (1990), based on inversion of *P*-wave polarities, azimuths to the stations, and the takeoff angles. We used only the events with at least 10 polarities and kept solutions constructed without any inconsistent polarities. Overall, the aftershock mechanisms are widely different, showing combinations of strike-slip, thrust, and very few normal solutions, as commonly observed in aftershock sequences. Only the northwestern termination of the rupture seems to show a predominant type of mechanisms, namely north–south-oriented normal faulting.

Local Earthquake Tomography (LET) of the Source Zone

We used our high-quality phase pickings and earthquake catalog for tomographic inversion. For this analysis, we applied the double-difference tomography method developed



Figure 7. (a) Magnitudes of aftershocks versus time starting from mainshock using data from the National Seismological Network (NSS). M_1 is local magnitude defined by the NSS. (b) The same information presented in part (a) with data from our temporary network (gray).

by Zhang and Thurber (2003). This method uses both the absolute and relative arrival times in a joint solution for event locations and velocity structure. It allows the production of a more accurate velocity structure in the region near the sources and, simultaneously, relative locations with a quality equivalent to the one obtained from the hypoDD method (Waldhauser and Ellsworth, 2000). As compared to previous LET methods, this improvement is particularly useful for our study, where, due to the field difficulties, all seismic stations are situated on one side of the fault.

From the nearly 1700 events located, we kept about 1100 events, which met restrictive criteria, that is, a number of readings exceeding 14, including at least two *S* arrivals, and a root mean square (rms) lower than 0.25 sec (Fig. 10). We preferred to use a small number of high-quality events, as more numerous poorly located events will introduce noise in the data and degrade the results. We obtained about 20,000 absolute travel times (~50% each *P* and *S* waves) and constructed from these 110,000 differential travel times for event pairs with interevent distances of less than 10 km. This distance choice is determined by the error values in the routine event location of our dataset (see the error diagrams in Fig. 6c).

The inversion grid finally chosen after routine tests is presented in Figure 11a in map view. It is rotated 42° clockwise, allowing the X nodes to be almost parallel to the coseismic fault trace. The distance between the nodes is, from the surface down to the maximal depth of the hypocenters, 4 km in the X direction and 3 km and in the Y direction, except in the central part along the fault, where the high density of events allowed a 2 km grid spacing. The initial velocity model is interpolated from the 1D model obtained through VELEST (Table 2) and used for the hypocentral location. After 20 iterations, the weighted rms travel-time residual was reduced from 0.23 to 0.07 sec.



Figure 8. Difference in epicenter positions seen during the summer 2004 experiment, between the local network (stars) and three other regional or worldwide networks (the circle represents the National Earthquake Information Center [NEIC], the square represents the Russian Academy of Sciences [RAS] Geophysical Survey Obninsk [OBN], and the triangle represents the Siberian Branch RAS Geophysical Survey, Novosibirsk [NVS]). N is the number of epicenter; D (in kilometers) is the average distance.



Figure 9. Calculated focal mechanisms for 69 aftershocks during the 2004 experiment built with at least 10 polarities.

The results of the tomography are presented in two figures. Relocated hypocenters are shown with magnitude on map view (Fig. 10a) along a longitudinal cross section following the coseismic fault trace (N132° E strike) (Fig. 10b) and along eight 4 km wide cross sections perpendicular to the trace (cross sections A-H projected N42° E; Fig. 10c). The P velocities are presented on Figure 11 where derivative weighted sum (DWS) values are higher than 10. The higher the DWS values, the better resolved the image; thus, the 100 and 1000 DWS-value contours are also drawn as markers of the final model resolution. The 3D velocity model presents a rather similar velocity zoning from the surface down to 9 km. As the surface layer is only resolved where the seismic stations are situated, we show on map view (Fig. 11a) the best resolved layer at a 3 km depth, which is representative of the shallower upper-crustal structures. The very poor resolution of the surface layer is clearly seen in Figure 11b. At 3 km depth (Fig. 11a), the mean P velocity is 5.83 km/sec. For discussion, we draw the coseismic rupture trace observed at the surface on the maps (Figs. 10a and 11a).

Comparing Figures 6a and 10a is not straightforward, as they present initial and final locations of different sets of events. Nevertheless, the spatial coincidence between the relocated aftershocks and the fault trace become clearer. In particular, the separation of the rupture into two branches in the southeast seems to be underlined by the aftershocks' alignments. In contrast, note that no aftershock remains under the east–west small (secondary) breaks near the Chuya river (Fig. 10a). The dense seismic swarm southwest of the fault (at about 50° N, 87.6° E) is still observed and more narrow. It is clear (Fig. 11a) that lower values of V_p (4.5–5.5 km/sec) follow the axial line of the source zone all along the main seismic fault system. A low-velocity strip follows the NCF



Figure 10. (a) Relocated aftershocks, together with coseismic rupture traces. Black lines are locations of cross sections plotted in (b) and (c), with shaded zones corresponding to the projected 4 km wide swath. Stars are the positions of the main aftershocks from ISC (see Fig. 2). (b) Along average rupture strike cross section including a complete set of aftershocks. (c) Cross-strike aftershock sections. Black arrows locate surface ruptures.

rupture almost continuously. In the southeastern region, where the fault system splays into two nearly parallel branches, two low-velocity oval strips oriented south-southeast are observed. Thus, at a depth of 3 km, the strict correspondence between the seismic fault and the narrow linear strip of low V_p values is observed. At a regional scale, the velocities may be linked to larger structures. The Kurai basin is characterized by lower velocities, more marked on its northeastern side. We may also identify low velocities under the westernmost part of the large Chuya basin, when the



Figure 11. (a) Map of P velocity at a depth of 3 km. Inversion grid used in tomography is presented in gray, inversion nodes being at the intersection of perpendicular lines. Shaded zones are not resolved; white lines are 10, 100, and 1000 contour values of iso-DWS from outside to inside. Black arrow-ended segments correspond to the location of cross sections plotted in part (b). (b) Cross sections through the P velocity model; the letters refer to lines in part (a).

Sukkor block located between both basins corresponds to higher velocities. The Kurai range as well as the Chuya range correspond roughly to higher velocity regions. For easier description, we will consider Figures 10c and 11b together. The *A* profile is located at the northwestern end of the rupture and of the aftershock cloud. The seismicity is

mainly concentrated between both fault traces, from the surface down to 13 km. A low-velocity vertical strip $(V_p = 4.3-5.5 \text{ km/sec})$ is observed down to the maximum resolved depth underneath the main rupture trace. Under the east-west secondary breaks north of the main rupture, one can identify a less contrasted velocity zone dipping towards the northeast. On the *B* profile, aftershocks are vertically aligned under the main rupture down to 13 km and are associated with lower V_p . On average, we observe the same disposition of velocity structures as on the *A* profile.

Going more to the southeast, on the C profile, the location of the aftershocks under the main rupture is more scattered. The events are concentrated under the fault down to about 7 km; then they occupy a larger volume with hypocentral depths reaching 18 km. An extremely concentrated cluster is observed to the southwest. All events are located within a vertical column, with a less than 4 km wide diameter, down to a depth of 12 km. No clear velocity anomaly corresponds to this cluster. As on profile B, we see no seismicity associated to the northernmost secondary ruptures. Away from the main fault, we still observe strips of low velocities dipping to the northeast on the C profile as well as on the D. On the D profile, the events are scattered, mainly to the southwest of the fault, in a wide and thick zone. Although less scattered on profile E, the aftershocks do not define a clear fault plane. On both sections F and G, the seismic activity becomes weaker and weaker. It is mainly concentrated under the rupture and is quasivertical. However, as far as the F profile, the tomographic models are mainly characterized by deep low-velocity strips dipping to the northeast. Finally, the H profile crosses the rupture where it splays into two branches. Aftershocks are aligned under both branches, from 3 to 13 km in depth, and associated to slightly contrasted low V_p zones.

Discussion

Very few data are available on the structure of the crust under the Russian Altay. No specific geophysical investigations were carried out in this region during the last few decades. Kabannik (2004) determined a velocity model of the upper half of the crust, down to a depth of 20 km. By his own method, he inverted the travel times of 291 earthquakes that occurred in 2002-2003 in the Kurai-Chuya region that were recorded by 32 seismic stations, 19 being local and the others being regional stations, on a $10 \times 10 \times 2$ km grid. Kabannik's results are difficult to compare with ours, as they are essentially less resolved and very rough. Emanov et al. (2004) is more directly comparable with our results, as the authors apply the same LET method to a completely independent set of data. They inverted the travel times of 407 aftershocks recorded during the months following the mainshock by temporary and permanent Siberian stations. Their starting velocity model was much faster in the upper crust, and initial event locations were quite different. The relocated aftershock cloud presents nearly the same elongated shape, but it is

much more scattered and shifted towards the northeast relative to the surface rupture trace. A more recent seismological study (Ulziibat, 2006) is based on a set of aftershocks, registered from a few days after the mainshock during several months, with three local stations and distant stations in Mongolia. Although the uncertainties in location are much larger, the main dimensions of the aftershock cloud, 80 km in the direction of the average rupture strike and with a depth of 20 km, are comparable to our results (Fig. 10b). Moreover, Ulziibat's source modeling of the mainshock and the largest aftershocks, along with the corresponding focal mechanisms, corroborated our result of a very steep fault plane slightly dipping to the southwest, implying mostly strikeslip right-lateral displacement together with a small reverse component.

The aftershock cloud shows more dispersion at the transition between fault sections I and II (Fig. 10a, and cross sections C and D in Fig. 10c). At a more detailed level, one can see several alignments of events striking at high angle to the general rupture direction in map view (Fig. 10a). We note that this is close to where the mainshock is located (Figs 2, 3, and 10a) and also where the surface rupture trace steps to the right. Our observations may suggest complexity in the fault plane at the transition between the North Chuya range and the Sukkor block (Fig. 3). It is interesting to note that the best fault model derived from InSAR (Nissen *et al.*, 2007) is also segmented into three parts, almost matching our sections I, II, and III.

The relocation of aftershocks recorded during the summer of 2004 allowed us to identify a column-shaped swarm southwest of the main rupture (Fig. 10). The aftershock column is as large as 4 km in diameter and extends from the surface to a depth of about 12 km (section *C*, Fig. 10c). It should be noted that most of these events occurred in less than 10 days during the summer 2004 survey and that this feature was not visible just after the mainshock (Emanov *et al.*, 2004; Ulziibat, 2006). We have no explanation for the shape of this peculiar sequence, because it is difficult to link to slip on a simple fault plane. This could be related to fluid migration in the upper crust even if no temporal migration is observed.

We note that the surface breaks that follow the Chuya riverbed west of Kurai city are not associated with aftershocks after relocation (Figs. 6a and 10a). While it has been suggested that these breaks were produced by one of the large aftershocks (Rogozhin *et al.*, 2007), it remains unclear whether these breaks are primary ruptures on well-identified faults or whether they are only ground fissures or slope breaks that do not root on a fault plane. The fact that no aftershocks are localized in this area, which is also a result obtained by Emanov *et al.* (2004) and Ulziibat (2006) for the three months following the mainshock, favors an interpretation that these breaks are secondary ruptures, possibly being triggered slip from the mainshock. However, the *P* velocity tomography shows north or northeast dipping structures beneath the surface breaks that may underline dipping reverse faults (Fig. 11b).

A clear outcome of the tomography is evidence of a long, narrow, low-velocity zone associated with the surface rupture trace (Fig. 11a). This zone is visible almost all along the rupture down to a depth of 17 km (Figs. 11a and 11b). A similar low-velocity zone has been shown along the San Andreas fault, however, principally in its creeping section (Dorbath *et al.*, 1996). Although we cannot exclude that the North Chuya fault is creeping, the existence of this low-velocity zone is indicative of a preexisting fault. This result corroborates field evidence of paleoseismic ruptures (Rogozhin *et al.*, 1998, 2003) despite the sparse geomorphic evidence of large accumulation of displacement along this fault.

Our results provide a robust determination of the seismogenic depth in Gorny Altay. The distribution of the aftershocks' cloud down to a depth of 17 km (Fig. 10b) indicates that the crust behaves similarly as in Mongolia where the seismogenic depth is taken to be around 20 km (Schlupp, 1996). In addition, if we take the aftershock cloud as representative of the rupture plane (approximately a rectangle of 17×80 km), then an average coseismic slip of about 2 m is needed to reach a magnitude M_w 7.2 (Kanamori, 1977). This means that the maximum slip is probably larger (3-4 m), although no offsets this large have been observed in the field. Larger slip, however, has been determined by correlation of optical Satellite Pour l'Observation de la Terre images (maximum slip of 4 m; Ulziibat, 2006; Barbot et al., 2008). We note that the characteristics of the event, the length of surface rupture trace of 60 km, a maximum slip of 4 m, and an average slip of 1-2 m all provide magnitude estimates of 7.2-7.3 following standard scaling relationships for strikeslip events (Wells and Coppersmith, 1994). On one hand, the larger slip values from the image correlation may be due to underestimating the vertical component of thrusting. On the other hand, the smaller slip values found in the field may partly result from errors in assessing the total coseismic offset when the rupture is characterized by en echelon fissures and subparallel branching. Similar difficulties in assessing coseismic offset have been encountered for even larger events, such as the Kokoxili event of 2001 in Tibet, where maximum slip was first determined to be around 16 m (Lin et al., 2002) and then revised to about half this value (Xu et al., 2006).

Field evidence of previous events along the same fault trace are clearly seen at places along reactivated fault scarps (Fig. 4). Large old landslides of unknown ages that may be of seismic origin are also seen in the region, in particular, near the largest landslide that occurred in 2003 (Rogozhin *et al.*, 2003, 2007). Large cumulative displacements are, however, not seen along the North Chuya fault, neither in the field nor from inspection of satellite images. Paleoseismic evidence (Rogozhin *et al.*, 2003, 2007) of a natural exposure across a pressure ridge indicates the occurrence of at least two events in the last 1000 yr, with the possibility of three events in the last 5000 yr, that is, on average, an earthquake every 500 to

2000 yrs considering characteristic behavior (Schwartz and Coppersmith, 1984). Given these numbers and the fact that the area was fully glaciated 16 ka and probably largely occupied by large glaciers until 12 ka, cumulative displacements may range from about 10 to 50 m. That no such large displacements are observed in the geomorphology suggests that the average recurrence time of large events is closer to 2000 yr. Given average coseismic displacements of 2 m, the maximum slip rate along the fault is thus 1 mm/yr or less, which is in rough agreement with geodetic global positioning system strain models that involve between 3 and 8 mm/yr of northeast–southwest shortening accommodated across the whole Altay belt (e.g., Calais *et al.*, 2006).

Conclusion

The Altay $M_{\rm w}$ 7.2 earthquake is the largest seismic event in recent time that occurred in Gorny Altay. Both the geological fieldwork and seismological studies in response to this event bring new insights about the seismic activity of this part of the Altay massif and about the crustal structure. The relocated aftershocks extend to a depth of about 17 km (Fig. 10), a depth also reached by the low-velocity zone under the surface rupture (Fig. 11), indicating that the seismogenic layer in Gorny Altay reaches almost 20 km (see also Barbot et al., 2008), a result similar to the one obtained for the crust in Mongolia from the study of the largest historical events (e.g., Schlupp, 1996). The dimension of the rupture plane $(80 \times 17 \text{ km})$ and an average slip of 2 m provide a magnitude estimate consistent with the moment magnitude determined from seismological and space geodetic modeling (M_w 7.2, e.g., Ulziibat, 2006; Nissen et al., 2007; Barbot et al., 2008; Harvard moment tensors). The slightly smaller surface displacements measured in the field in the central section of the event (e.g., Rogozhin et al., 2003) may result from broad deformation that is not recognizable in the field and/or difficulties in assessing total slip where the surface rupture shows complicated geometry. The surface rupture only extends for about 60 km, a length somewhat shorter than the extension of aftershocks at both extremities of the mapped ruptures.

Slip rate of the North Chuya fault is most probably smaller than 1 mm/yr, with the recurrence time of large M 7 earthquakes being likely greater than 2000 yr. The difficulties to identify such fault prior to an event is here, in Gorny Altay, certainly linked both to its slow movement as well as to the glacial and periglacial environment of the region. As a consequence, active faults, which have clear geomorphological expressions, should thus focus our attention, such as the Kurai fault, for example, that makes up a clear cut along the southern Kurai range (Fig. 3).

Data and Resources

Seismograms used in this study were collected using a mobile network of the Schmidt Institute of Physics of the Earth (Russian Academy of Sciences) and belong to this Institute.

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