Influence of station topography and Moho depth on the mislocation vectors for the Kyrgyz Broadband Seismic Network (KNET)

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SUMMARY

Deviations of slowness and backazimuth from theoretically calculated values, the so-called mislocation vectors, are measured for the Kyrgyz Broadband Seismic Network (KNET) in the Tien Shan region. 870 events have been analysed for arrivals of P and PKP waves from all azimuths. The deviations of slowness and backazimuth show a strong trend with values up to 1 s deg−1 for slowness values for waves arriving from the North and South and backazimuth deviations of, in some cases, more than 10° for waves arriving from the East and West. Calculating the traveltime deviations of the stations for topography of the Tien Shan region and Moho depth values appropriate for this area shows that most slowness and backazimuth deviations can be reduced to very small values. The remaining mislocation vectors show no strong trends and are on average smaller than 0.2 s deg−1 for slowness and 2° for backazimuth values, which is within the error bars of these measurements. Results from array methods that rely on the knowledge of the backazimuth values show much improved resolution after the correction of the mislocation vectors which shows the importance of knowing and correcting for structures directly beneath arrays.

Key words: Time series analysis; Body waves; Site effects.

INTRODUCTION

Since the advent of digital seismometers, many seismic arrays have been deployed in recent years. Several of those are designed to study shallower structures (e.g. the currently installed USArray, Meltzer et al. (1999)) but are also often used for studies of the deep Earth (e.g. Weber 1993; Weber & Wicks 1996; Koper et al. 2003; Rost & Thomas 2009) and many new array methods have been developed since the introduction of arrays (see e.g. Rost & Thomas (2002, 2009), Schweitzer et al. (2002), Rost & Garnero (2004) for reviews on array techniques).

One of the main application of seismic arrays is the ability to measure both slowness and backazimuth of incident signals (Frosch & Green 1966; Whiteway 1966; Filson 1975; Nawab et al. 1985; Suteau-Henson 1990; Rost & Thomas 2002, 2009), which helps to identify phases or distinguish between arrivals. Before being detected by seismic stations a seismic wave passes through heterogeneities in the Earth’s interior and can be reflected, refracted and scattered. These reflected, refracted and scattered waves can have slowness vectors (i.e. backazimuth, the angle from North to the great circle path measured at the station, and horizontal slowness, the incidence angle to the surface) that are different to that of the first arrival, that is, the direct P wave. Array seismology methods are based on determining the angle of incidence (slowness) of a wave at the array, and in most cases it is assumed that the theoretical backazimuth is the correct backazimuth (e.g. Krüger & Weber 1992; Rost & Thomas 2002). Another assumption is the plane wave approximation: for most array techniques the wave is considered as plane wave, which for teleseismic waves and small aperture arrays is a valid assumption.

Before reaching the array, the seismic wave also has to pass through the underground beneath the array, which may be characterised by local heterogeneities. It has been shown that these heterogeneities can affect the calculation of slowness and backazimuth at an array (e.g. Niazi 1966; Berteussen 1974, 1976; Faber et al. 1986; Krüger & Weber 1992; Koch & Kradolfer 1997, 1999; Bondár et al. 1999) and lead to slightly different values compared with the predicted values for slowness and backazimuth from global Earth models such as IASP91 (Kennett & Engdahl 1991) or ak135 (Kennett et al. 1995). Slowness and backazimuth values are calculated through the measurement of differential traveltimes of the wave arriving at the stations of an array, the so-called delay times (e.g. Rost & Thomas 2002). The deviations from the theoretical values of slowness and backazimuth are called mislocation vectors and in case of larger structures beneath the array are often consistent for different arrival directions (e.g. Glover & Alexander 1969; Engdahl

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& Felix 1971; Berteussen 1974, 1976; Faber et al. 1986; Krüger & Weber 1992; Bondár et al. 1999). The origin of slowness and backazimuth anomalies, that is, the mislocation vectors, measured at seismic arrays often lies directly underneath the array in the crust, or lithosphere (e.g. Berteussen 1976; Faber et al. 1986).

Since for several array-processing methods the knowledge of the true slowness or backazimuth of an incoming wave is essential, purely taking the theoretically obtained values from the knowledge of source and receiver locations, without taking account of the mislocation vectors, may lead to misleading results (Rost & Thomas 2002). Producing a vespagram (slowness stack) (Davies et al. 1971) with a backazimuth that differs from the true backazimuth of the seismic arrival can shift the resulting slowness values by large amounts and even make seismic arrivals non-detectable (Rost & Thomas 2002). It is therefore essential to know the mislocation vectors for a seismic array if array methods are used.

In this study we are determining the mislocation vectors for an array in the Tien Shan region, the Kyrgyz Broadband Seismic Network (KNET) (Vernon 1994, 1998; Mellors 1995) that has been running continuously for 20 years and has been used for several array studies in the past (e.g. Mellors 1995; Mellors et al. 1997; Bump & Sheehan 1998; Ghose et al. 1998; Wolfe & Vernon 1998; Martynov et al. 1999, 2004; Xu et al. 2006; Rost & Thomas 2009; Thomas & Billen 2009; Landes et al. 2010; He & Wen 2011).

TIEN-SHAN AND KNET

The Tien Shan region located in Kyrgyzstan is one of the most prominent mountain belts in central Asia (Burtman 1975) and a good example for contemporary intracontinental orogenesis (Tapponnier & Molnar 1979; Zhiwei et al. 2009). It was formed in the past 50 million years, mainly caused by the collision of the Indian with the Eurasian Plate and underthrusting of the Indian Plate underneath the Eurasian Plate (Burtman 1975; Molnar & Tapponnier 1975) and the shortening rate across the Tien Shan reaches values of up to 20 mm yr

\(^{-1}\) (Abdrahmatov et al. 1996; Reiger et al. 2001). The Tien Shan region consists of a system of East–West trending mountain ranges with a length of some 2500 km, reaching from the Kyzyl Kum in the West to the Gobi Desert in the East and a width of up to 400 km (Tapponnier & Molnar 1979; Roecker et al. 1993; Ghose et al. 1998, Zhiwei et al. 2009). The region is seismically active (Molnar & Tapponnier 1975; Tapponnier & Molnar 1979; Ghose et al. 1998), and since 1910 six major earthquakes occurred with magnitudes of more than 7.0 (Ni1978). Since 1991 the Tien Shan region is monitored by the KNET, a network of digital broadband three-component seismometers (Vernon 1994; Mellors 1995; Ghose et al. 1998), installed and operated by Scripps Institution of Oceanography, the Kyrgyz Institute of Seismology and the Institution of the Russian Academy of Sciences Joint Institute for High Temperatures RAS.

The KNET became operational in 1991 September (Vernon 1994, 1998; Mellors 1995). The Network covers part of the transition zone between the northern Tien Shan mountains and the Kazakh Platform, a number of major tectonic features are monitored by the network including several thrust faults in the Tien Shan and the Talasssa-Fergana fault (Molnar & Tapponnier 1975; Tapponnier & Molnar 1979; Vernon 1994; Mellors 1995; Ghose et al. 1998). Due to the mountainous area, the stations are spread over a large altitude range from 655 m (Station CHM) up to 3850 m (Station UCH). Fig. 1 shows the location of the KNET in central Asia.

DATA AND PROCESSING

The data used to measure the mislocation vectors for the KNET are recorded in the period from 1996 June to 2010 April. We select teleseismic events in the distance range from 20° to 90° as well as events in the distance range from 150° to 170°. Over 1000 events are analysed, 870 of them are studied in detail. The events must fit the following three criterions (i) they must lie in the selected distance range: 20° to 90° for P waves, 150° to 170° for PKP-wave (to avoid contamination from the interference of PKP waves near the caustic at 144°), (ii) The P arrival and the PKP arrival, respectively must be visible on at least seven seismogram traces including station KBK. The magnitude as well as the depth of the events are not important, thus the magnitude ranges from \(m_b = 4.7\) to \(m_b = 8.6\) and the depth ranges from 0 to 670 km. A map of all events is shown in Fig. 2.

We mainly analyse the unfiltered data, yet some seismograms are filtered using the filter wssn-sp. We conducted several tests using a range of filters to verify that filtering has no effect on the mislocation vectors. We do not apply further corrections such as static correction (Sheriff 1968; Capon 1973; Harjes & Henger 1973; Ingate 1966; Green et al. 1966; Capon 1973; Harjes & Henger 1973; Ingate et al. 1985; Navab et al. 1985; Suteau-Henson 1990). In this study we calculate slowness and backazimuth directly from the observed relative traveltimes, that is, the delay times, a method that is implemented...
The measured values are then compared with the theoretical values determined by using the model IASP91 (Kennett & Engdahl 1991) for the given source–receiver combination. The resulting mislocation vectors are then displayed in a stereographic diagram in a similar way to Krüger & Weber (1992). The vectors point from the observed slowness/backazimuth pair (Obs) to the theoretical value (PDE). The theoretical slowness is calculated with the TauP Toolkit (Crotwell et al. 1999) and the theoretical backazimuth is calculated from trigonometrical aspects, assuming a spherical Earth. For these calculations the knowledge of latitude and longitude of every event is necessary which is taken from the NEIC Preliminary Determinations of Epicentres (PDE).

RESULTS

The measured full mislocation vectors for the KNET are shown in Fig. 3(a) for all earthquakes used in this study. For a better understanding of the two individual components, we have also split the mislocation vectors up in their slowness component in Fig. 3(b) and the backazimuth component in Fig. 3(c), respectively, following the analysis in Krüger & Weber (1992). In Fig. 3(d) the slowness- and backazimuth anomalies $\delta_s$ and $\delta_{\text{baz}}$ are plotted against the theoretical backazimuth, including the average deviation calculated with the program Seismic Handler.

In the three figures, Figs 3(a)–(c), a strong trend is visible: All mislocation vectors for all arrival directions point to the South. This means that for events in the backazimuth range from 0° to 180° a smaller backazimuth and in the range from 180° to 360° a larger backazimuth is measured compared to the theoretically calculated values. The pattern for the slowness measurements is similar: for events coming from the north (backazimuth from 270° to 90°) measurements of the slowness result in values that are too large compared with the theoretically calculated values. For events arriving from the south (backazimuth from 90° to 270°) slowness values smaller than expected are measured. The maximum slowness anomaly is up to 0.93 s deg$^{-1}$ and for backazimuth measurements in some cases more than 10°.

Fig. 3(d) shows the anomalies with corresponding error bars. The errors for both slowness and backazimuth measurements are smaller than the variations and a clear trend for both slowness and backazimuth is visible. In the backazimuth range around 135°, 220° and 270° to 300° some large backazimuth deviations of up to 30° are visible, that correspond to the PKP-arrivals. Although the associated mislocation vectors have average length they result in higher backazimuth deviations because of the smaller slowness. In Fig. 4 we show the traveltime residuals measured for each station, where KBK is the reference stations for which all residuals are 0.

CORRECTIONS OF MISLOCATION MEASUREMENTS

As Berteussen (1976) discussed, the origins of mislocation vectors at NORSAR are sited at the receiver side and located in the upper 60 km of the crust underneath the array stations. Other studies found similar results (e.g. Krüger & Weber 1992; Bondár et al. 1999). Also the array topography can affect the measurement of slowness and backazimuth (e.g. Bokelmann 1995a). The array topography is expected to have a strong influence on the KNET mislocation vectors since the maximum altitude difference of the array stations amounts to 3195 m. This topography directly affects the relative traveltimes and therefore the measurement of slowness and
backazimuth because the localization depends on the correct measure of the relative traveltimes. Another important effect is a variable depth of the Moho discontinuity that changes significantly underneath the KNET array stations (e.g. Bump & Sheehan 1998; Vinnik et al. 2004, 2006; Steffen et al. 2011). The influence on the relative traveltimes comes from the strong velocity change at the Moho discontinuity. A seismic wave crossing a deep Moho travels a longer path through lower velocity material and thus needs more time compared to a wave crossing the Moho at shallow depth, travelling a shorter distance through the upper crust. Furthermore a dipping discontinuity leads to a reorientation of the propagation vector of an incoming wave. In the following sections we will look at the effects in more detail.

ARRAY TOPOGRAPHY

The high altitude differences of the KNET array stations strongly and systematically distort the relative traveltimes, compared to a flat topography array configuration. This is because the localization
method assumes a flat topography, which leads to results deviating from the PDE-solution. Three of the four northernmost KNET array stations USP, CHM and TKM are situated lower than 1000 m while the three southernmost stations UCH, AML and KZA are placed on heights exceeding 3000 m. With the station KBK (1760 m) as reference station this implies that a signal at southern stations is recorded later than expected and earlier than expected at the northern stations. Assuming a seismic velocity of 5.8 km s\(^{-1}\) for the upper crust (Kennett et al. 1995) the maximum altitude difference of 3195 m between the two Stations CHM (655 m) and UCH (3850 m) implies traveltime residuals of more than 0.5 s that cannot be neglected. The effect of the topography on the mislocation vectors is explained schematically in Fig. 5. Here the traveltimes calculated for a flat earth are given in Fig. 5(b), but with topography, the traveltimes for station 1 (St. 1) which is lower than the central station (St. 0) is shorter whereas for the higher station 2 (St. 2) the traveltimes is longer. To compensate for these traveltim differences, the wave has an apparent backazimuth, that is, in this case, further from the North, that is, larger.

The effect of the station topography on the relative traveltimes can be calculated from the array station heights, a seismic velocity in the upper crust and the theoretical slowness to take into account the angle of incidence. We took a seismic velocity of 5.8 km s\(^{-1}\) for the upper crust, which represents the uppermost crustal layer in several 1-D Earth models (e.g. Kennett & Engdahl 1991). Investigations of the crustal structure of the Kyrgyz Tien Shan give upper crust velocities (e.g. Ghose et al. 1998; Lei & Zhao 2007; Omuralieva et al. 2009; Lei 2011). However, we have chosen the velocity from the model IASP91 because the above studies show slightly differing results and also reveal small-scale velocity structures that we are not considering here.

With this travelttime correction the influence of the station topography on the mislocation vectors can be eliminated. The result of the topography correction is presented in Fig. 6. The corrected
mislocation vectors are smaller than the uncorrected ones shown in Fig. 3 but there is still a small trend visible and it seems that topography is not the only factor affecting the mislocation vectors.

It is important to mention that the topography effect can be already eliminated by applying a static correction (Sheriff 1968; Turhan Taner et al. 1974). However since we were processing only uncorrected data to determine the mislocation vectors, the topography still has a strong influence on the data. The effect of station topography can also be included in the calculation of traveltimes (e.g. Schweitzer et al. 2002).

TOPOGRAPHY OF THE MOHO

Several studies detect a significant change in Moho thickness beneath the Tien-Shan region (e.g. Belousov et al. 1980; Kosarev et al. 1993; Cotton & Avouac 1994; Bump & Sheehan 1998; Vinnik et al. 2004, 2006; Shin et al. 2007; Makarov et al. 2010; Bagdassarov et al. 2011; Steffen et al. 2011). A variation of depth of the Moho discontinuity under seismic arrays can influence the mislocation vectors (e.g. Niazi 1966; Otsuka 1966a,b; Sheppard 1967; Glover & Alexander 1969; Greenfield & Sheppard 1969; Iyer & Healy 1972; Berteussen 1974, 1975; Krüger & Weber 1992). Two reasons are usually given, which are both based on the strong velocity increase at the Moho with increasing depth: (i) Seismic waves crossing the Moho in different depths cover different distances through the crust and due to the lower seismic velocity in the crust will be delayed or advanced relative to each other, which systematically affects the relative traveltimes. (ii) Seismic waves will change their propagation direction due to a dipping Moho according to Snell’s law. Here we consider both effects. For the calculation of the traveltime residuals we only consider the depth of the Moho directly underneath each station. For the estimation of the dependency of the slowness vector reorientation on the theoretical slowness vector and the dipping discontinuity we follow the approach given in Niazi (1966).

We use the Moho depth from Steffen et al. (2011) and assume a flat Moho with a NEE strike and a dip of about 6° directly underneath the array stations. We estimate the P-wave velocity change at the Moho from studies of Xu et al. (2002, 2007), Lei & Zhao (2007) and Zhiwei et al. (2009) to be 6.7 km s$^{-1}$ above and 7.75 km s$^{-1}$ below the Moho. The velocities given in those studies are lower compared with the global average, probably due to the upwelling of hot mantle material (Roecker et al. 1993; Xu et al. 2002, 2007; Lei & Zhao 2007; Lei 2011). The remaining mislocation vectors after applying both Moho corrections together with the array topography correction (cf. Fig. 6) are shown in Fig. 7. The slowness and backazimuth deviations are much smaller compared with uncorrected data shown in Fig. 3 and look more random. Yet there are still some areas that show systematic but small trends in the mislocation vectors, for example, in the backazimuth range from North to 60° where the observed mislocations seem to have been overcompensated by our correction.

DISCUSSION

In the previous section we have discussed two causes for the major part of the observed slowness- and backazimuth anomalies: the station topography as a major effect and the Moho topography as a minor effect. In Fig. 8 the histograms of the backazimuth- and slowness residuals before and after the corrections are represented. Even though the corrections improve the measurements of slowness and backazimuth, there are still small slowness- and backazimuth anomalies remaining. The remaining mislocation vectors do not show a systematic trend as the vectors in Fig. 3. Yet there are backazimuth ranges that show systematic deviations, for example, in the range from 350° to 60° or from 150° to 210°. These systematic backazimuth deviations could possibly be explained by local
heterogeneities such as low velocity zones or even anisotropy and it may be useful to investigate also small-scale structures that have some effect on the relative traveltimes or cause a reorientation of the waves.

Several studies report the existence of low velocity zones underneath the Kyrgyz Tien Shan region (e.g. Ghose et al. 1998; Lei & Zhao 2007; Xu et al. 2007; Zhiwei et al. 2007, 2009; Omuralieva et al. 2009; Lei 2011), which are presumed to be due to hot rising mantle material. Those studies show comparable results: (i) strong and very local P-wave velocity anomalies; (ii) smaller or at least equal crustal velocities in the southern Kyrgyz Tien Shan compared with the northern Kyrgyz Tien Shan. These velocity anomalies can affect the relative traveltimes and in further consequence the mislocation vectors. The large-scale small velocity difference from North to South is expected to have a small influence on the mislocation vectors and we did not examine its effect further. The strong and small-scale low velocity regions, however, could influence the relative traveltimes of a few stations only and could lead to backazimuth dependent deviations of the mislocation vectors. Since the models of low velocity regions and their locations differ we decided not to use these to correct for the models since each would produce small corrections on the data but they do not necessarily explain all the data completely. Correcting all observed mislocation vectors would produce a very small-scale regional model and is beyond the scope of this study.

The study of Bokelmann (1995b) showed that local anisotropy in the vicinity of the array stations can affect the wave propagation direction and contribute to the mislocation vectors. There exist few studies that show the presence of anisotropy in the central Tien Shan region (e.g. Wolfe & Vernon 1998; Martynov et al. 2004; Vinnik et al. 2007; Zhiwei et al. 2007). We expect this influence to be small because of the larger aperture of KNET, and since the few existing studies show different results we chose not to take anisotropy into account.

For the calculation of slowness and backazimuth from the delay times, the plane wave approximation is generally used. By restricting our events to teleseismic distances, we ensure that the assumption holds. If, however, signals generated in the vicinity of the array would be used (e.g. scattering) the waves could not be assumed plane (for an example see Thomas et al. 1999). Another assumption used in this study is that seismic waves are coherent across the array. For smaller arrays such as the KNET or GRF array the coherency is given for teleseismic P waves and longer period waves. For increasing array aperture the coherency of the signal could decrease and before stacking data or correlation measurements the coherency should be tested. A similar problem poses the superposition of waves and a subsequent loss of coherency of a wave across the array. This could happen in the case of PKP waves near the caustic for an array of several degrees aperture.

To show the importance of the mislocation vectors we present two vespagrams for two different events recorded at KNET. For each event we produce two vespagrams: taking the theoretical backazimuth (PDE-Solution) and for the second vespagram using the observed backazimuth. The vespagrams are shown in Fig. 9. The Figs 9(a) and (b) show the vespagrams of an event in New Zealand with a theoretical backazimuth $\Theta = 130.86^\circ$ in a distance of 120° to investigate PP-precursors (underside reflections at discontinuities in the upper mantle and transition zone). In Fig. 9(a) the $PP$-wave as well as the $PKiKP$ phase are clearly visible while the precursors to $PP$ are not distinguishable from the noise. In Fig. 9(b) we use the observed backazimuth of 120° and here the P410P precursor...
Figure 9. Four 4th-root vespagrams of two events recorded at KNET. (a) Event in New Zealand, 2007 September 30 09:47:52, \( m_b 6.6 \) distance: 120.07 \(^\circ\), backazimuth \( \theta = 130.68^\circ \), depth 31 km). The vespagram is produced with the theoretical backazimuth \( \theta = 130.68^\circ \). The PP- as well as the PKiKP arrivals are visible. A P410P arrival cannot be distinguished from the surrounding noise. (b) Vespagram for the same event but with the observed backazimuth of 122\(^\circ\). The phases PP and PKiKP are clearly visible and now the P410P phase can be detected at the correct traveltime. (c) Event in Indonesia (2003 May 5 23:04:45, \( m_b 6.2 \), distance 60.9\(^\circ\), backazimuth 114.19\(^\circ\), depth 56 km). The vespagram was produced with the theoretical backazimuth of 114.19\(^\circ\). The P-arrival is visible but its slowness is not easily measurable. In the time range from 13 to 17 s the pP-arrival is expected but not visible. (d) Slowness-Vespagram for the same event as (c) but with the measured backazimuth of 109\(^\circ\). The slowness of the P-arrival is now easier to be determined and in the time range from 13 to 17 s a clear arrival is visible that likely represents the pP phase. The focussing of other phases is better and the slowness values can be determined more easily.

CONCLUSION

We have analysed 870 events recorded at KNET to determine the deviations of slowness and backazimuth of seismic waves from the theoretically calculated. These mislocation vectors show a strong trend with all vectors pointing to the south. The backazimuth deviations vary in a range of up to 15\(^\circ\) whereas the slowness deviations vary between \( \pm 1 \) s deg\(^{-1}\). To find the main contributors to the slowness and backazimuth deviations we correct the traveltimes of P and PKiKP by applying these corrections the size of the deviations is reduced considerably and the mislocation vectors show no obvious trends any more. Small-scale structures could perhaps explain the remainder of the deviations. The importance of considering the true backazimuth of a seismic wave is shown in a few examples of events recorded at KNET where, after using array methods with the observed azimuth instead of the theoretical azimuth, small phases become visible in vespagrams and can be analysed.

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Figure 10. Example for a detection of a small event ($m_b$ of 4.7) in a distance of approximately 61°. The data are filtered with the wwssn_sp filter. (a) Single trace of station KBK. (b) Stacked traces of all KNET stations after beamforming using the theoretical azimuth of 121° (top panel) and the corrected backazimuth (114°, bottom panel). The slowness value used for both beams is 6.3 s deg⁻¹. (c) Vespagram with the theoretical backazimuth (121°). (d) Vespagram with a backazimuth of 114°. The black arrows in (c) and (d) indicate the theoretical slowness value and traveltime for this event. The focussing of the event is improved when using the corrected value and the slowness agrees with the theoretical value.

REFERENCES


1995), SRTM data are downloaded from International Centre for Tropical Agriculture (CIAT) (Reuter *et al.* 2007; Jarvis *et al.* 2008).

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Meltzer, A. et al., 1999. The USArray initiative, GSA today, 9, 8–10.


