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Seismic anisotropy beneath the Northern Apennines (Italy) and its tectonic implications

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Abstract. We examined shear wave splitting in *SKS* and *S* phases from 22 teleseisms at 10 temporary stations on a transect across the Northern Apenninic arc. The array, near 43°N, spans from Corsica Island across the Tyrrhenian region and the Apenninic belt to the Adriatic coast. We applied particle motion, covariance matrix decomposition, and cross correlation methods to estimate the polarization direction of the fast split - shear wave (ϕ) and the delay time between split phases (δt). Most of the analyzed shear waves show clear evidence of splitting. The ϕ in the Adriatic foreland and in the Apennines are approximately parallel to the strike of the mountain belt (NW-SE). The largest δt correspond to the highest elevations, suggesting that anisotropy is related to the compressional tectonics which built the Apennines, and that this tectonic compression involved at least the entire lithosphere. In the Tyrrhenian area we observe ϕ oriented about E-W, suggesting a reorientation of the mantle fabric due to astenospheric flow, responsible for the E-W post-orogenic extension observed at the surface.

Introduction

The Apennines are part of a complex collisional plate boundary in the central Mediterranean, between Africa and Eurasia [*Malinverno and Ryan*, 1986; *Royden et al.*, 1987, and references therein]. The Northern Apennines arc formed by building up several NE - verging thrust sheets with eastward progressive migration of the compression front (upper Miocene-Pleistocene). In the Tyrrhenian region, post-orogenic extension has been accompanied by thinning of the crust, normal faulting, ductile deformation, volcanic activity, and high heat flow (Miocene in Corsica to Plio-Pleistocene in Tuscany) [*Jolivet et al.*, 1994]. Afterwards, extensional processes affected the Apennines (Plio-Pleistocene), where graben-like structures trending N-S to NW-SE overprint the compressional features. On the Adriatic margin, in contrast, the outer front of the belt probably is still in compression [*Patacca and Scandone*, 1987] (Fig. 1).

Several models have been proposed for the geodynamic evolution of the Apennines, and there is still debate on whether the formation of the mountain range is related to westward subduction of the Adriatic plate or not. Here we address this question using seismic measurements of anisotropy at upper mantle depths obtained with a temporary seismic array spanning from Corsica, to the Adriatic coast. Recent tomographic studies give some clues about upper mantle structure beneath the Northern Apennines revealing the existence of a high velocity body in the upper 200 km [*Cimini and Amato*, 1993]. This observation and the presence of subcrustal seismicity in this area [*Selvaggi and Amato*, 1992] suggest the existence of south-westward subduction of the Adriatic lithosphere underneath Tuscany.

Our study is part of a multidisciplinary project supported by the European Community, with the goal of providing new data to constrain better the geodynamic evolution of the Apennines. The seismological studies are mainly focused on investigating the deep structure. Three temporary seismic transects across the Apenninic belt were planned, two of which have already occurred (Fig. 1). The instruments used are digital seismographs (RefTek) recording continuously and equipped with three component 5-s (Lennartz and Kinemetrics) or broad-band (Guralp) sensors; the response at low frequencies of 5-s sensors (normalized response at 10 s is about 0.2) forced us to analyze only data from large earthquakes. Teleseisms recorded constitute a data base for tomographic inversions, receiver function studies and shear wave splitting analyses. This paper presents the first results of our shear wave splitting study using the data collected during the 1994 Northern Apennines transect. It reveals the existence of significant anisotropy beneath the Apennines. Because anisotropy is related to upper mantle and crustal deformation induced by past and present tectonic events [*Silver and Chan*, 1991, hereafter S&C] the characterization of seismic anisotropy along the transects provides new keys to understanding the geodynamic development of the Apenninic belt. The close spacing of our stations gives us a rare opportunity to monitor variations in splitting parameters between different tectonic environments.

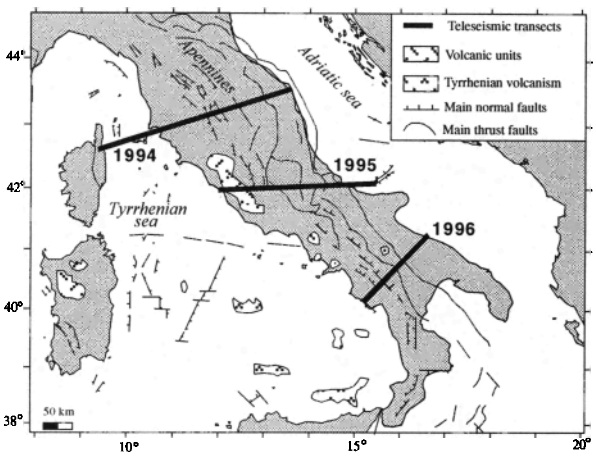

Figure 1. Simplified structural map of Italy with the approximate location of the GeoModAp teleseismic transects: 1994 (Northern Apennines), 1995 (Central Apennines), and 1996 (proposed Southern Apennines)

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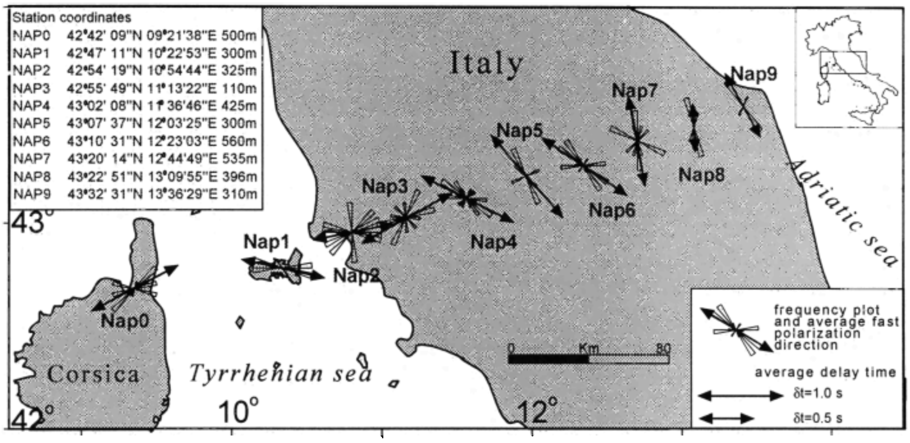

Figure 2. Map of Northern Apennines transect and observed fast polarization directions, ϕ, using the covariance matrix decomposition method. Arrows are average ϕ and their lengths are proportional to the mean delay times, δt. The rose diagrams are frequency plots of the ϕ measurements normalized to each station's number of data.

SKS and *S* splitting analysis

During the 1994 Northern Apennines transect we deployed 10 stations from Corsica Island to the Adriatic coastline (Fig. 2). The transect is 340 km long, and the average station spacing is about 30 km. For our analysis we first selected 8 events with magnitudes > 5.7 and epicentral distance $\Delta > 86°$; this subset of data contains *SKS* phases with a good signal-to-noise ratio (Table 1). The use of *SKS* is preferred in splitting analyses because this phase is sensitive only to the receiver side anisotropy and has a theoretically known polarization direction coincident with the back-azimuth of the earthquake [*S&C*]. We added to the study teleseismic S phases (Table 1) to enlarge our data set and because their polarization directions improved the azimuthal coverage. Events with Δ of $\approx 85°$ are characterized by SKS and S arrivals close in time; we considered the S wave since this phase is more energetic at these distances [*Bullen*, 1963]. We analyzed waves with periods of $\approx$ 10 s which corresponds, in the uppermost mantle, to a wavelength of $\approx$ 45 km; considering these wavelengths and the short crustal raypaths the influence of a crust 25 - 35 km thick is marginal. We visually inspected all the seismograms and the particle motion of the horizontal components rotated into radial and transverse directions choosing a time window suitable for splitting analysis. We applied covariance matrix decomposition and cross correlation to evaluate fast polarization directions (ϕ) and delay times (δt) (Fig. 3). These methods are described in detail by *Zhang and Schwartz* [1994]; here we adapted this method to analyze teleseismic data. In covariance matrix decomposition we subdivide a 20 s window centered at the shear wave arrival, into 40 windows of 1 s duration (40 samples each) with overlap of 0.5 s. For each time window we calculate the rectilinearity R

$$R=1-((\lambda_2+\lambda_3)/2\lambda_1)^n$$

the shear angle S (dip of the polarization vector)

$$S=\cos^{-1}(u_Z)$$

and the polarization angle Φ

$$\Phi=\tan^{-1}[(u_{ew}\,\mathrm{sign}(1,u_Z))/(u_{ns}\,\mathrm{sign}(1,u_Z))]$$

where λ_1, λ_2, and λ_3 are the three eigenvalues of the covariance matrix, u_Z, u_{ns} and u_{ew} are the normalized eigenvectors for the largest eigenvalue (λ_1), and n=1/3, this value of n is introduced to amplify the variability range of *R*. The sign fortran function resolves the 180° ambiguity by taking the positive vertical component of u_Z [Jurkevics, 1988]. The Φ corresponding to the shear wave onset (identified by high rectilinearity and shear angle close to 90°) coincides with the ϕ. The δt was determined by cross correlating the two split waves (Fig. 3).

In order to check our observations we applied an alternative method, cross-correlating the same 20 s window of the horizontal components after successive rotations in steps of 1°. By using a grid search across rotation angle and lag we found the azimuth and the lag corresponding to the largest cross-correlation coefficient: these are our estimates of ϕ and δt. We classified results as "excellent", "good", "marginal"

Table 1. Analyzed events

Date 1994	Or.Time* (hhmmss.ss)	Δ (°) *	α (°) *	Depth (Km)*	m_b *	Phase used
06/05	010930.15	86.97	59.06	11.	6.1	SKS
06/06	204740.53	86.61	270.81	12.	6.4	SKS
06/09	003316.23	91.85	252.89	631.	7.0	SKS
06/20	090902.91	35.37	99.80	9.	5.9	S
06/29	182233.58	63.02	69.65	10.	5.9	S
06/30	092321.35	45.19	77.70	227.	6.1	S
07/04	213641.96	93.47	293.96	15.	6.1	SKS
07/21	183631.74	79.85	40.30	471.	6.5	S
07/31	051539.58	30.35	97.89	43.	5.3	S
08/02	141754.87	80.57	20.27	170.	5.8	S
08/14	004622.49	85.06	28.57	33.	5.9	S
08/14	013114.84	85.00	28.63	33.	6.1	S
08/18	044259.68	85.06	28.53	33.	6.1	S
08/19	100251.84	97.90	240.96	565.	6.4	SKS
08/20	043851.66	84.70	29.17	33.	6.1	S
08/21	155601.80	62.87	36.66	33.	5.7	S
08/22	172638.20	141.55	42.90	148.	6.1	SKS
08/28	183719.70	85.03	28.51	14.	6.0	S
08/29	173620.62	51.49	221.37	10.	5.4	S
08/30	061336.16	85.02	28.53	54.	6.1	S
09/13	042801.08	88.23	50,61	37.	5.8	SKS
09/28	163952.21	100.00	88.57	643.	5.7	SKS

*NEIC (USGS) bulletins

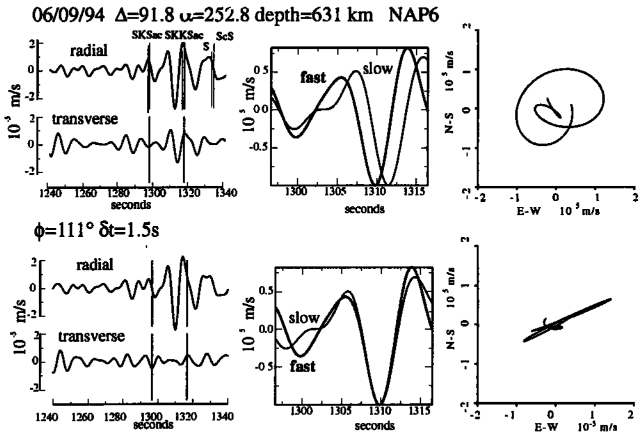

Figure 3. Example of splitting. Top: *SKS* radial and transverse component seismograms, filtered between 0.02 and 0.2 Hz, fast and slow wave and particle motion in the NS EW plane before the removal of the δt on the slow wave; bottom: after removal of δt. The polarization direction is now consistent with the back-azimuth of the event.

(weight 3, 2, 1 respectively, [*Evans et al.*, 1995]) according to the cross correlation coefficients, to the degree of linearization in particle motion after removal of the lag, and, for the *SKS*, according to the coincidence between polarization direction and event back-azimuth. Therefore, at each station, we evaluated the directional weighted average of ϕ and δt and their standard deviations (Table 2). We compared values resulting from the two different methods and we found, for most of the stations, except NAP3, good agreement in the average ϕ: differences <20° are consistent with anticipated uncertainties in polarization direction [*Zollo and Bernard*, 1989]. We rely on the results of the covariance matrix decomposition method in what follows and in figures 2 and 4. The analyzed *S* and *SKS* waves show evidence of splitting at all sites, some sites with significant variability in ϕ for different events. Delay times, δt, of about 1 s or larger are observed consistently at most stations.

Results and Discussion: tectonic implications

At the stations on the outer front of the Apennines (NAP8, NAP9, Fig. 2 and 4) the ϕ are almost parallel to the strike of the main thrust fault (NW-SE), in agreement with the results of other studies in collisional zones [*Vinnik et al.*, 1992, *Barruol and Souriau*, 1995]. The average δt at these stations are about 0.5 s. On the mountain belt (NAP7 to NAP4) the ϕ slightly rotate from NNW in the East toward WNW in the West with δt > 1 s. In the Tyrrhenian region (NAP3 to NAP0) the ϕ are ENE and E-W, the δt are 0.5 - 1.0 seconds. Generally the observed δt are much larger than what is estimated for the crust; at NAP6, analyzing local earthquakes we find ϕ consistent with the one from teleseisms and δt less than 0.1 s confirming that crustal contribution to anisotropy is marginal.

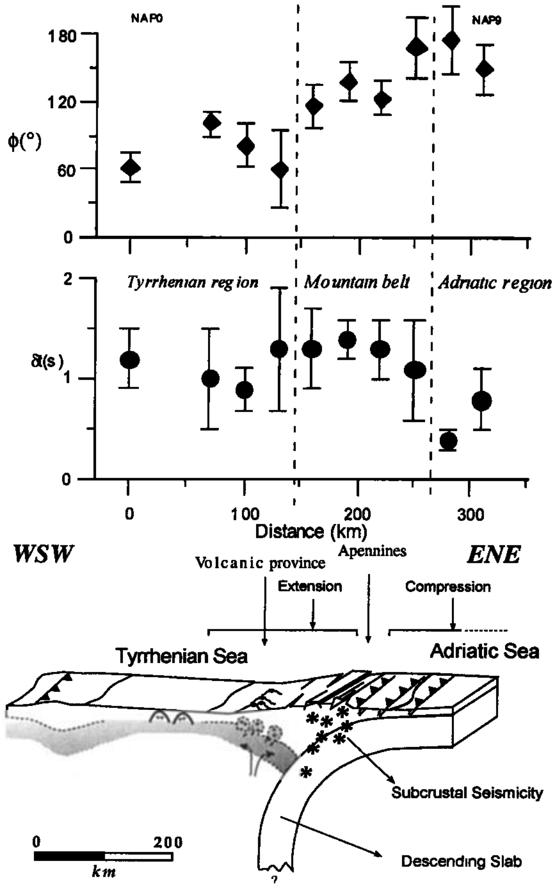

Figure 4. Schematic model showing the hypothesized geometry of Adriatic subducted lithosphere [modified after *Cimini and Amato*, 1993]. Vertical and horizontal scale are the same. Average ϕ and δt, calculated by the covariance matrix decomposition method, with their standard deviations, are plotted above.

Table 2. Average Anisotropic Parameters and Errors

		Cov. Matrix Decomposition				*Cross Correlation Analysis*			
Station	*#meas.*	*ϕ (°)*	*S.D. ϕ (°)*	*δt (s)*	*S.D. δt(s)*	*ϕ (°)*	*S.D. ϕ (°)*	*δt (s)*	*S.D.δt(s)*
NAP0	9	61.8	13.5	1.2	0.3	80.6	12.7	1.1	0.5
NAP1	4	100.9	9.6	1.0	0.5	88.7	18.5	0.4	0.3
NAP2	6	81.5	18.6	0.9	0.2	73.6	22.1	0.6	0.2
NAP3	6	59.9	34.6	1.3	0.6	112.3	18.3	0.7	0.4
NAP4	8	116.6	19.2	1.3	0.4	104.6	21.2	1.2	0.5
NAP5	3	138.9	16.8	1.4	0.2	122.2	7.9	0.8	0.3
NAP6	8	123.4	15.3	1.3	0.3	139.1	23.3	0.9	0.3
NAP7	6	169.5	26.6	1.1	0.5	159.5	27.6	1.0	0.8
NAP8	2	176.5	30.8	0.4	0.1	166.0	3.5	0.3	0.1
NAP9	2	151.0	22.7	0.8	0.3	131.7	4.2	0.4	0.3

S.D.= standard deviation

The three regions with distinct anisotropic parameters correspond to the weakly deformed Adriatic foreland, the highly deformed mountain belt and the recently active Tyrrhenian volcanic area (Fig. 4). The ϕ seem to be related to the most recent significant episode of deformation in each region. The ϕ suggest that the orogenesis of the Apennines caused a preferential alignment of upper mantle anisotropic minerals parallel to the belt axis. This NW-SE ϕ are detectable from NAP5 eastward, in the region where the compression is very recent or still active. This region corresponds to the area underlain by the Adriatic lithosphere (Fig. 4). Conversely, the Tyrrhenian area, after the build-up of the mountain range, underwent widespread EW extension. Thus the E-W ϕ observed can be related to this post-orogenic episode of deformation.

An estimate of the thickness of the anisotropic region may be made from the size of the delay time δt. Assuming an average anisotropy of about 4 - 6 % a δt of 1 s corresponds to an anisotropic layer about 115 km thick [*Mainprice and Silver*, 1993]. Hence, observed δt suggest that the anisotropic layer thickness ranges between 35 km on the Adriatic side, 160 km beneath the Apennines and 100 km in the Tyrrhenian area. Of course our estimates are based on the previous assumptions of similar intrinsic anisotropy across the whole transect. Indeed the main part of the anisotropy probably is in the upper 100-150 km because of the small scale variability of ϕ and δt and because of the relation of the anisotropy to the surface geology and to the stress in the brittle crust as inferred from earthquakes focal mechanisms and borehole breakouts [*Montone et al.*, 1995]. It is interesting to note that the extent of the area affected by along axis fast polarization direction (NAP5 - NAP9) is larger than that of the active compressional region in the crust (external front of the belt) suggesting the presence of Adriatic lithosphere beneath the mountain belt.

The trend of increasing δt observed from station NAP9 to NAP5 can be related to the increased lithospheric thickness [*Barruol and Souriau*, 1995; *Helffrich*, 1995] and to the possible lower intrinsic anisotropy in the weakly deformed region. In the Apennines the increased lithospheric thickness may be due to the present configuration of subduction or collision, where the Adriatic plate is bent beneath the belt (Fig. 4). The δt at Tyrrhenian stations appears to require a contribution from the asthenosphere since lithospheric thickness there is roughly 40 km [*Panza*, 1984]. This thin lithosphere cannot explain delays of about 1 s unless we hypothesize an intrinsic anisotropy much larger than 4 to 6%, which would be surprising. Further more, it is interesting to note that anisotropy found using Pn phases [*Mele et al.*, 1995] are consistent with the ones we found in the Adriatic lithosphere but not in the Tyrrhenian one suggesting for this latter region an asthenospheric source of anisotropy. The E-W direction of ϕ observed in the region between Corsica and Tuscany is consistent with the direction of extension observed form surface geology [*Jolivet*, 1994], possibly suggesting that the anisotropy here is related to recent astenospheric flow, which in turn control the extension in the lithosphere.

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