Introduction. In this study we present some examples of measurement and modeling of the earthquake cycle, in different tectonic domains; we discuss some common problems concerning geodetic data interpretation and fault modeling related to the availability of geological data at the surface and at depth.

The study of surface deformation is one of the most important topics to improve the knowledge of the deep mechanisms governing the seismic cycle itself and, eventually, improve the assessment of seismic hazard. To measure the crustal ground deformation associated to fault activity and to the earthquake-cycle we use DInSAR (Differential Interferometric Synthetic Aperture Radar) (Massonnet and Feigl, 1995), MAI (Multi Aperture Interferometry) (Scheiber and Moreira, 2000), (Bechor and Zebker, 2006) and multitemporal InSAR methods (Ferretti et al., 2001; Berardino et al., 2002; Hooper, 2007). Currently these techniques allow measuring short-term ground displacement with centimetric accuracy (DInSAR and MAI), and ground velocities with an accuracy of up to one millimeter per year over time periods of several years (Casu et al., 2006). These levels of accuracy make interferometric methods suitable for the study of tectonic processes, typically affected by deformation rates ranging from millimeters to centimeters per year. Furthermore, these methods allow measurement of ground movements occurring on the different time scales of the earthquake cycle, from the nearly instantaneous deformation caused by seismic dislocations (Massonnet et al., 1993), to the slow strains of the interseismic phase (Wright et al., 2001).

Many conceptual, numerical, analytical and analog models of the earthquake cycle have been proposed to explain seismological, geological, geomorphological and geodetic data.
It is today accepted that the seismic cycle can be subdivided in three main different phases: interseismic, coseismic and postseismic (e.g. Scholz and Kato, 1978). Actually, the interseismic phase can be again subdivided into a purely interseismic step and a preseismic one, but the state of knowledge concerning the latter is still vague (e.g. Deng et al., 1998).

The post-seismic deformation occurs soon after the seismic event and it can be subdivided in two phases, characterized by short and long-term deformation. The short-term deformation can usually be attributed to afterslip and/or pore pressure readjustments, which take place over periods ranging from a few hours to a few months after the earthquake. Deformation cannot be accurately measured by interferometric methodologies if the image sampling interval over the area is too long (several days or months). The long-term postseismic deformation is instead related to the viscoelastic relaxation that occurs in the lower crust and upper mantle, following several months or years (depending on the magnitude) after the earthquake (e.g. Segall, 2002). This kind of deformation, also called viscoelastic rebound, is difficult to be isolated using InSAR data, because: a) it is often characterized by ground velocities at the lower boundary of the InSAR measurement capability (<1 mm/yr); b) it is spread over long distances and can be confused with long-wavelength InSAR error sources as well as with interseismic deformation. In this work we do not consider crustal deformation caused by the viscoelastic rebound because of the small magnitude of the considered earthquakes and the small length of the data time-series; we focus our attention on the interseismic, coseismic and postseismic (afterslip and/or pore pressure readjustments) phases.

From the geological point of view the earthquake cycle reveals itself through field evidence, such as abrupt offsets or diffuse deformation of lithological reference layers, fluviatile or marine terraces, depositional or erosional landforms, faults escarpments, etc. The seismic cycle and the fault activity are also studied using paleo-seismological trenches, where geologists can measure and date stratigraphic layers to evaluate long term averages of strain rates and displacements. Starting from geological or paleoseismological field data, it is possible to evaluate the mean slip rate along a fault, and recognize the main seismic events (those which have ruptured the surface); these slip rates are usually averaged over a time span of thousands or tens of thousands of years (Pantosti et al., 1993).

To study the seismic cycle in the long term (interseismic phase), we need to integrate geological and geodetic data, which entails integrating slip rates averaged over many seismic cycles, and present day ground velocity maps. To reconcile these different data we need to identify the sources responsible for the present day strain accumulation (and geodetic velocity), and possibly know the long term slip rate along them. In this way we can compare the geologic and the geodetic slip rates. The geodetic slip rate can only be obtained by appropriate modeling of the geodetic data; normal methods for interseismic geodetic data inversion neglect transient deformation processes and estimate slip rates by assuming slip on a fault to occur by steady creep only below a locking depth, in an elastic half-space, over the course of the earthquake cycle. On the other hand, for short-term studies of the earthquake cycle (co and post seismic phases), is very important to understand and identify the seismic sources and the interaction between each other. In fact, during a seismic crisis, typically several different data are available to the researchers concerning the hypocenters, preexisting geological and ad hoc field data, cinematic of the sources, magnitudes, punctual displacement data (GPS) and displacement fields (SAR); one of the main goals is to reconcile all of these data to define the seismic sources (geometries, kinematics and locations) and to investigate the relationship between the main sources and the static stress variations along the modeled fault planes (e.g. Pezzo et al., 2013).

In this work, we present the results of the multitemporal InSAR-SBAS analysis (Berardino et al., 2002) for the measurement of low interseismic ground velocities, in particular we study the interseismic deformation in the Gargano promontory (Southern Italy) and in the Doruneh region (Northern Iran) (Pezzo et al., 2012). In both cases we focused our attention (i.e. the modeling) on the most prominent tectonic structures of the areas, the Mattinata Fault, for the
Fig. 1 – East (A) and up (B) components mean velocity maps of the Gargano area with main tectonic features. In panel A, the blue colors represent positive velocities (eastward) and in red the negative ones (westward). In panel B, the blue colors represent positive velocities (upward) and in red the negative ones (downward). Black flags represent the GPS permanent stations and the red line is the trace of the velocity profiles (2.5 km buffered) shown below. In panel C, D and E the mean LOS velocity maps from (C) ascending track 156, (D) descending track 206 and (E) descending track 435 (from Pezzo et al., 2012). Positive velocity values (blue colors) indicate ground movement toward the satellite along the LOS direction (inclined 23° from the vertical), negative ones (red colors) indicate the opposit. In red we show the DFS system trace; other faults in the region are in black. Black triangles indicate the reference point; black boxes mark the ~20 km buffered velocity profiles reported with the profile (A), (B) and (C), used in the data inversion; dashed black lines mark the intersection with the DFS trace. We show velocity error bars of 1 mmy⁻¹. All the maps are geo-referenced in UTM WGS84, Zone 40 N.
Gargano, and the Doruneh Fault System for Northern Iran. Both test sites are characterized by prevalent transcurrent tectonic regimes with low deformation rates (few mm per yr).

We also present some examples of coseismic studies in which the geological field data availability or absence played a fundamental role in the definition of the seismic sources: the 2012 Emilia seismic sequence and the 2008 Baluchistan (Central Pakistan) seismic swarm. For both interseismic and coseismic datasets we performed the data inversions using the elastic model of Okada (1985), with the aim to investigate its applicability in the general context of the study of the earthquake-cycle in different tectonic domains, analyzing the limits of the method and the contribution of surface and deep geological data.

Concerning the postseismic deformation, we present results concerning two strong earthquakes affecting the Italian peninsula in the last four years: the 2009 L’Aquila and the 2012 Emilia earthquakes. In this cases we show the fundamental role of geological field data concerning the survey of the afterslip ground faulting with respect to gravitational phenomena (for the L’Aquila event), and the survey of the hydrogeological dynamic and paleo-topography analysis of the epicentral area to discern the tectonic contributions to the ground deformation (for the Emilia events).

Results. Ground deformation during the interseismic and preseismic phase. We used multitemporal InSAR techniques for the measurement of low interseismic ground velocities, in particular we studied the interseismic deformation in the Gargano promontory (Southern Italy) and in the Doruneh region (Northern Iran).

The Gargano Promontory is an ENE-WSW oriented topographical and structural high representing a portion of the Apulian foreland extending into the Adriatic Sea. This area is characterized by active (instrumental and historical) seismicity that can originate strong earthquakes reaching intensities of XI MCS. The E-W Mattinata Fault (MF) has been frequently proposed as a natural candidate to represent the major seismogenic structure in the Gargano area; however the spatial distribution of the recorded seismicity does not concentrate along this structure. Using the SBAS multitemporal DInSAR technique (Berardino et al., 2002) we analyzed a SAR dataset of 88 descending (from 1992 to 2010) and 46 ascending (from 1995 to 2008) ERS and ENVISAT images to retrieve high resolution, mean velocity maps of the Gargano area (Southern Italy). Combining ascending and descending line of sight velocity maps, we obtained the East and Up component velocity maps (Figs. 1A and 1B). The Up velocity component (Fig. 1B) shows no vertical movement in the inner Gargano, whereas, in the East velocity component (Fig. 1A), we identify a NW-SE oriented maximum gradient of about 0.03 mm yr$^{-1}$ per km in the inner Gargano, across the MF, in good agreement with GPS data and the main stress axis orientation. We also evaluate a deformation rate of about 40-50 nanostrains yr$^{-1}$ for the East component in the inner Gargano along E-W oriented velocity profile. These values are higher than the regional ones calculated using GPS, borehole or seismological punctual data. We modeled the MF using the elastic dislocation model proposed by Okada (1985). Our measurements are compatible with a right-lateral kinematics of MF with a slip rate of about 1.5 mm yr$^{-1}$ to a depth of 15 km, as shown by comparing observed and modeled velocity profile (Profile A of Fig. 1A). Our results confirm the slip rate values proposed by many authors, and the right-lateral geodynamical interpretation of the MF as a fundamental active junction in the dynamic equilibrium between the different motion of the Adria and Apulia microplates. We interpreted the interferometric signal and we inverted the SAR maps using the MF source parameters, subsequently evaluating the inversion goodness by comparing our results with bibliographic data. Without fundamental a priori information about the geometry of the fault, the hypocenter distribution and geological knowledge of the area, it would have been impossible to carry out a correct data interpretation and source modeling.

Concerning the Northern Iran site, we used the SBAS DInSAR analysis technique to estimate the interseismic deformation along the western part of the Doruneh fault system (DFS) (Pezzo et al., 2012). We processed 90 ENVISAT images from four different frames from
ascending and descending orbits (Figs. 1C, 1D and 1E). Using a simple dislocation approach we modeled 2-D velocity profiles obtaining a good fit to the observations (Profile A, B and C of Figs. 1C, 1D and 1E). Our model confirms the general left lateral kinematics of the DFS, but an additional important thrust component along the WFZ is necessary to fit the observations (rake angle of 34°±4°). The modeled rake angle results in the left-lateral component being about 2/3 of the slip rate, which is in agreement with the long-term record as reconstructed by structural and geomorphic observations. A steep fault dip to the North (~60°) is well constrained by the observations, and is in agreement with field observations on the Western and Central DFS. Our modeled slip rate of 5±1 mm yr\(^{-1}\) is the first quantitative estimate of strain accumulation for the Western DFS, corresponding to ~4 mm yr\(^{-1}\) of pure horizontal movement. In the long term, Fattahi \textit{et al.} (2007) estimates ~2.4 mm yr\(^{-1}\) of left lateral slip rate on the central fault zone, by Infrared Stimulated Luminescence Dating (ISLD) of one Holocene alluvial fan. In our model, the anticorrelation between slip and rake indicates that some vertical component (rake > 0) is needed to explain the observations, otherwise the slip rates for pure strike slip fault would be unrealistic compared to the geological estimates. None of the latter are available for the Quaternary uplift rate along the western fault zone (WFZ). Extrapolating the vertical component of our modeled slip rate (2.4±1 mm yr\(^{-1}\)) in the geological past does not seem to justify the relatively low relief across the WFZ (500-700 m), but a number of different factors, e.g. observation and/or model uncertainties, spatio-temporal slip rate variations, interaction with close tectonic structures could justify this discrepancy.

Based on a geologically determined average slip per event of 4.7 m, the recurrence interval along the Central DFS has been estimated to be ~2000 yr (Fattahi \textit{et al.}, 2007). The slip rate we estimated for the WFZ would require a slip per event of ~10 m to obtain the same recurrence time. This is unrealistic, especially if we consider the field observations and the segmentation model by Farbod \textit{et al.} (2011). Accepting a similar slip per event for the WFZ and the CFZ, including a conservative uncertainty estimate of 20%, we obtain a recurrence interval varying between 630 and 1400 yr. The lack of strong seismicity in the last 1500 yr along the WFZ may be due to the incompleteness of the historical seismic catalogues (Ambraseys and Melville 1982), and our results show that recurrence intervals below 1000 yr cannot be excluded.

\textbf{Ground deformation during the Coseismic phase.} Regarding the contribution of geological constrains to the interpretation and modeling of coseismic displacement SAR maps, we present two examples in which a priori geological information had a very different significance.

The first example concerns two earthquakes occurred in the Emilia region, Northern Italy, respectively on May 20\(^{\text{th}}\) 2012, M\(_L\) 5.9, and May 29\(^{\text{th}}\), M\(_L\) 5.8, inverting COSMO-SkyMed and Radarsat-1 surface displacements and GPS observations (Fig. 2A) (Pezzo \textit{et al.}, 2013). The Emilia seismic sequence filled a seismic gap existing at least since the year 1000 A.D. (Rovida \textit{et al.}, 2011), and might therefore give important information on the mechanisms of strain accumulation and release in this area. Thus, some considerations can be made based on our results about the long-term strain accumulation. The location of the coseismic deformation measured by InSAR data corresponds to part of the Mirandola and Ferrara folds, located, under the Po alluvial plain. This evidence supports the long-term geomorphic analyses that attribute to the growth of the same folds the wide northward bend of the Po river course and the deviation of the Secchia and Panaro rivers (Burrato \textit{et al.}, 2003). Both the 20 and 29 May 2012 sources were found to be well modeled by -E-W, S-dipping thrust faults with a flat-ramp geometry (Figs. 2C and 2D), corresponding to the Mirandola and Ferrara thrusts. Furthermore, we identified a displacement pattern of ~10 cm towards the satellite sensors, not associated to any of the largest aftershocks (Fig. 2B). The pattern is temporally co-located or following the first event and preceding the second one. Spatially it is located halfway between the displacement fields of the two main events. We investigate some possible interpretations of our observations, favoring the hypothesis of a slip along the fault plane of the May 29\(^{\text{th}}\) event, supported by the results of a Coloumb Failure Function analysis, which suggests an increasing
Fig. 2: (A) Cumulative displacement map spanning all main events of the sequence, obtained from the Radarsat-1 12 May–5 June interferogram. (B) Cumulative displacement from 12 to 27 May and 4 to 5 June, obtained by subtracting the COSMO-SkyMed 27 May–4 June displacement from the map shown in (A). Solid-line black rectangles indicate the footprint of the COSMO-SkyMed satellite image pairs. The insets in the top-right corners contain a zoom of the area enclosed by the dashed-line black rectangles. Arrows indicate the ground-projected line-of-sight (from ground to satellite) and the satellite flight path. (C) and (D) The slip distribution (1.5 X 1.5 km patches) along the 20 and 29 May sources, respectively. Purple spheres represent the hypocenter (http://iside.rm.ingv.it; M >2) relative to the following time spans: (C) 17–28 May 2012 and (D) 29 May–11 June 2012 (from Pezzo et al., 2013). Observed displacement maps from ENVISAT interferograms: (E) 06/05/2008–02/12/2008, (F) 04/03/2006–08/11/2008 and one MAI ENVISAT (G) 23/09/2008–02/12/2008 maps relating to the 28–28 October 2008 earthquakes. (H) Coseismic displacement map of the 09 December 2008 event from 02/12/2008–06/01/2009 interferogram. Satellite paths and line-of-sight (LOS) directions are shown in the boxes; black lines indicate the surface projections of the modeled faults. (I) and (J) the slip distribution (1.5 X 1.5 km patches) along the 28–29 October 2008 and 09 December sources, respectively. In panel B, in transparency the two mainshock fault planes.
load on the May 29th fault plane, following the mainshock. Two seismic sources were obtained using the fault plane parameters provided by several published geological studies in this area (e.g., Boccaletti et al., 2010; Picotti and Pazzaglia, 2008). This information is essential in this area, since, due to the symmetric shape of the deformation pattern and its small N-S component, the North/South dipping fault ambiguity could not be solved by SAR and GPS data inversion alone.

The second example concerns the three largest events of the 2008 Baluchistan (western Pakistan) seismic sequence, namely two Mw 6.4 events only 11 hours apart and an Mw 5.7 event 40 days later. We used Synthetic Aperture Radar Differential Interferometry (DInSAR) (Figs. 2E, 2F and 2H) and Multi-Aperture Interferometry (MAI) (Fig. 2G) to constrain the sources of these events. Our InSAR surface displacement maps and subsequent modeling results suggest the sources of the two main earthquakes of the Baluchistan 2008 seismic sequence were two NW-SE oriented right-lateral strike-slip faults (Figs. 2I and 2J). The modeled fault planes are found to be almost vertical and quasi parallel to each other, forming an about 90° angle with an ENE-WSW oriented fault responsible for the largest aftershock; the latter is a vertical left-lateral strike slip fault located between the two main sources. Moreover, CFF analysis suggests that the second mainshock fault plane was not overloaded by the first mainshock in spite of the very brief lapse of time between the two events. On the contrary, the December aftershock fault plane was intensely loaded by the occurrence of the October mainshocks.

These results are insightful when interpreted in the tectonic context of the Quetta syntaxis. In fact, the latter is placed in crucial junction between two blocks characterized by opposite relative motion, namely the northward motion of the Kirthar range and the southward motion of the Sulaiman Lobe. The former can be considered the right block of the left-lateral strike-slip Chaman fault system, whereas the latter is considered a transpressive zone in the northwestern part of the Indian subcontinent (Yadav et al., 2012), with a southward extrusion accommodated by the SE verging thrust fault surrounding the lobe itself and the left-lateral Kingri fault (Rowlands, 1978). In this complex tectonic context, our quasi parallel mainshock fault planes are in good agreement with a right-lateral shear zone located in the Quetta syntaxis. Moreover, the CFF suggests the October 28th and 29th earthquakes were two independent mainshocks characterized by a similar magnitude, mechanism and geometry. Conversely, concerning the December 9th aftershock, the stress increase along the fault plane due to the mainshocks suggests that it was likely triggered by the October earthquakes. Thus, at this scale, in the area included between the two mainshock fault planes, we can suppose a reorientation of the stress field due to the general right-lateral displacement of the blocks. Under this assumption, our observations, together with the modelled fault geometries, suggest that at a local scale this area could be affected by left-lateral shear zone, with a bookshelf type deformation, in a wider regional tectonic context of right-lateral shear zone, confirmed by our mainshock modelling.

This test case is characterized by the almost total absence of geological data at surface and at depth, as well as of any coseismic ground evidences of surface faulting. Constraints for source modeling are provided only by the abundance of SAR measurements (three independent motion components could be measured for the main events) and from seismology. In particular, the MAI technique was crucial in solving the fault plane ambiguity determined by moment tensors. Our results and hypothesis would however have to be confirmed by a field geological survey and a more accurate seismological study of the seismic sequence, as well as of stress readjustments and reorientations analysis during the seismic sequence evolution.

Ground deformation during the postseismic phase. The first postseismic deformation analysis we present concerns the well-known deformation following the Mw 6.3 L'Aquila earthquake occurred on 06/04/2009 (Lanari et al., 2010; D'Agostino et al., 2012).

We used 25 COSMO-SkyMed SAR images (beam 09 asc.) to obtain a postseismic deformation time series and mean velocity map (spanning 12/04/2009 to 13/10/2009) (Fig. 3A).
Fig. 3: (A) Postseismic mean velocity map of the 2009 L’Aquila earthquake area spanned from 12/04/2009 to 13/10/2009, obtained using 25 COSMO-SkyMed SAR images of the ascending beam 09. (B) Example of displacement time series evaluated close PAGA GPS station. (C) Postseismic mean velocity map of the 2012 Emilia seismic sequence area spanned from 20/07/2012 to 27/04/2013, obtained using 15 COSMO-SkyMed SAR images of the ascending beam b10h, superimposed to the structural model of Italy (Bigi et al., 1983) In A and B, positive velocity values (blue colors) indicate ground movement toward the satellite along the LOS direction (inclined 23° from the vertical), negative ones (red colors) indicate the opposite. (D) Example of displacement time series evaluated in the maximum line of sight displacement. (E) and (F) N-S displacement profile calculated across the 20/05/2012 earthquake epicentral area (E) and far from the entire seismic sequence (F); blue rhombuses are the postseismic displacement, whereas green triangles and red squares indicate the preseismic displacement obtained from ERS (from 30/09/1992 to 05/06/1999) and ENVISAT-ASAR images (from 15/01/2007 to 06/09/2010).
We also performed a correction of atmospheric effects correlated with topography, estimated from the SAR data itself in non-deforming areas. The time series analysis in the epicenter area show a typical exponential trend (Fig. 3B), so that we can mainly attribute the deformation pattern to the afterslip occurred along the 06/04/2009 earthquake fault plane. Based on ground based SAR interferometry, some authors have estimated the post seismic deformation along the fault plane trace to be of several centimeters (Wilkinson et al., 2012, 2010); despite the actual difficulty in field identification of centimetric surface deformation, it appears fundamental, soon after a strong earthquake, to provide a temporally continuous field deformation survey to identify the co and postseismic deforming areas. On the other hand, in addition to the afterslip correlated pattern, the postseismic velocity map shows many other features related to gravity, like deep seated gravitational slope deformations (DSGSD). Thus, a correct interpretation of the postseismic velocity map cannot be carried out without a detailed field investigation. Furthermore, the latter is crucial to correctly interpret possible transient deformation patterns from one tectonic structure to another, as well as the evolution of the seismic sequence (e.g. Pezzo et al., 2013).

As a second test-case, we present a post seismic multitemporal SAR analysis of the 2008 Emilia seismic sequence, in which 15 COSMO-SkyMed stripmap SAR images (Fig. 3C). Also in this case we observe a quasi-exponential afterslip deformation pattern in the epicentral area (Fig. 3D). In addition we detect a velocity pattern not directly attributable to the postseismic phenomena. In fact, local subsidence occurs due to fluid migration and compaction of the Po plain sediments. Moreover, there is a rough correspondence between the growth anticline and the positive (approach to the satellite) velocity values. Vice versa, negative velocity patterns correspond to the locations of the tectonic synclines. Furthermore, we observe a general agreement between the preseismic and postseismic, albeit for the afterslip pattern located in the epicentral area (Figs. 3E and 3F).

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