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SOVRAPPOSIZIONE DI DEFORMAZIONI
ESTENSIONALI E CONTRAZIONALI DURANTE
LO SVILUPPO DI PRISMI OROGENICI:
ESEMPI NELL'APPENNINO SETTENTRIONALE
E NEL SALAKH ARC (OMAN).

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ABSTRACT

The long term evolution of orogenic wedges is characterized by complex interactions between tectonics, gravitational body forces, and surface processes. It is influenced by changes of large-scale geodynamic conditions, geometrical configuration of plate margins, mechanical properties of the materials involved in the deformation, and climate conditions. Within this framework, contractional and extensional deformations may coexist in different sectors of a thrust wedge or cyclically affect the same area. In particular, extensional pulses developed during the contractional orogenic history of thrust wedges have been documented in various contexts and predicted by geomechanical models. Deformation structures associated with such extensional events contribute to the regional-scale evolution of orogens, influencing fluid circulation within the upper crust and modifying mechanical properties of thrust wedges. Therefore, sin-contractional extension have remarkable scientific and economic significance.

This Ph.D. thesis investigates extensional deformation structures developed during the orogenic contractional history of two thrust wedges: the Oman Mountains and the Northern Apennines. Research is done with a multidisciplinary approach, and close attention is paid to the palaeohydrological evolution of the deformation structures and their influence on fluid circulation through the upper crust.

The case study of the Jabal Qusaybah (Oman) allows investigating on an anticline developed at shallow structural levels, at the front of a thrust wedge. Despite this structural position, the evolution of the fold is characterized by interactions between contractional, extensional, and left-lateral strike-slip deformations. The Northern Apennines case studies involve km-scale extensional structures associated with the evolution of entire wedge sectors. In the Val di Lima case, a low-angle extensional fault system is unequivocally framed within the contractional orogenic history, being affected by superimposed folding due to out-of-sequence thrusting. In the Tellaro case, the extensional fault system cuts through the Punta Bianca anticline, being likely associated with the late orogenic thinning of the internal sector of the Northern Apennines. The Lunigiana case includes both the previous scenarios.

NOTES FOR THE READER

This manuscript is composed by different chapters, which are organized as follows.

1) RIASSUNTO ESTESO

Summary of the following chapters, written in Italian.

2) INTRODUCTION

Short introduction about orogenic wedges, sin-contractual extensional deformations, and their relevance.

3) OROGENIC WEDGE EVOLUTIONARY PROCESSES

Description of the accretionary wedges and characterization, through analogue modeling experiments, of the main processes that influence their long-term evolution.

4) JABAL QUSAYBAH ANTICLINE

Jabal Qusaybah anticline case study (Oman Mountains), that allows investigating the interactions between contractional and extensional deformations during the evolution of a single anticline, located at shallow structural levels at the front of an orogenic wedge.

5) VAL DI LIMA FAULT SYSTEM

Val di Lima fault system case study (Northern Apennines), that allows investigating large-scale structures developed in non-metamorphic conditions within an accretionary wedge, cyclically affected by compressional and extensional deformations.

6) TELLARO DETACHMENT

Tellaro Detachment case study (Northern Apennines), that allow investigating a late-orogenic large-scale low-angle extensional fault system characterizing the internal portion of an orogenic wedge.

7) LUNIGIANA FAULT SYSTEM

Lunigiana fault system case study (Northern Apennines), that allows investigating late-orogenic high-angle extensional structures characterizing the internal portion of a thrust wedge.

8) GENERAL CONCLUSIONS

Final chapter to highlight the general conclusions of the present work.

BIBLIOGRAPHY

Comprehensive list of references cited in previous chapters.

Research activities have been done within the framework of different projects, coadiuvated by international collaborations with academic and industrial partners. Analogue modeling experiments have been performed in the laboratory Géosciences Montpellier of the Université Montpellier II (France) under the supervision of Dr. Jacques Malavieille. Research on the Jabal Qusaybah anticline is part of a joint project between the University of Parma (Italy) and the Katholieke Universiteit Leuven (Belgium), funded by Shell Global Solutions. Research activities on paleofluid evolution have been done in collaboration with Prof. Rudy Swennen and Dr. Mahtab Mozafari (KU Leuven).

Research on the Val di Lima fault system and Tellaro Detachment case studies have been funded by the Italian Ministry of Education, University and Research (MIUR) through PRIN project “La deformazione nelle rocce carbonatiche: implicazioni per la caratterizzazione e la modellizzazione dei serbatoi naturali di geofluidi.” (PRIN 2009, National Coordinator E. Tondi, Local Coordinator F. Storti). Research activities on paleofluid evolution have been performed in collaboration with Prof. Philippe Muchez and Prof. Rudy Swennen (KU Leuven).

Research on the Lunigiana fault system has been supported by ENI Exploration & Production that allowed the consultation and interpretation of subsurface data in their offices in Milan (Italy). Seismic interpretation has been done in collaboration with Dr. Mirko Carlini and Dr. Andrea Artoni (University of Parma).

This manuscript is written in English to favor its circulation among the scientific community and to allow the publication of selected chapters as articles on specific scientific journals. Only the first chapter, that constitutes a summary of following contents, is written in Italian.

RIASSUNTO

L'evoluzione a lungo termine dei prismi di accrezione è caratterizzata da interazioni complesse tra tettonica, forze di volume gravitazionali, e processi superficiali. È influenzata da cambiamenti delle condizioni geodinamiche a grande scala, della configurazione geometrica dei margini di placca convergenti, delle proprietà meccaniche dei materiali coinvolti nella deformazione, e delle condizioni climatiche. In questo contesto, deformazioni contrazionali ed estensionali possono coesistere in porzioni diverse di uno stesso prisma o interessare ciclicamente la stessa area. Pulsi estensionali sviluppati durante la storia orogenica contrazionale di prismi di accrezione, in particolare, sono stati documentati in numerosi contesti e previsti dai modelli geo-meccanici. Le strutture deformative associate a tali pulsioni estensionali contribuiscono all'evoluzione a scala regionale degli orogeni, influenzando la circolazione dei fluidi nella crosta superiore e modificando le proprietà meccaniche dei prismi di accrezione. I pulsioni estensionali sin-contrazionali hanno perciò una notevole importanza scientifica ed economica.

Nell'ambito di questa Tesi di Dottorato vengono investigate strutture deformative sviluppate durante la storia orogenica contrazionale di due prismi di accrezione: le Montagne dell'Oman e l'Appennino Settentrionale. Le attività di ricerca sono caratterizzate da un approccio multidisciplinare; l'evoluzione paleoidrologica delle strutture e i loro effetti sulla circolazione dei fluidi nella crosta superiore sono investigati con particolare attenzione.

Il caso di studio di Jabal Qusaybah (Oman) offre la possibilità di studiare una struttura anticlinale sviluppata a livelli strutturali superficiali, al fronte del cuneo tettonico. Nonostante la sua posizione strutturale, l'evoluzione della piega è caratterizzata da interazioni tra deformazioni contrazionali, estensionali, e trascorrenti. I casi di studio nell'Appennino Settentrionale comprendono strutture estensionali chilometriche associate all'evoluzione di interi settori del prisma. Nel caso della Val di Lima, un sistema di faglie estensionali a basso angolo è univocamente inquadrato in un contesto contrazionale, essendo coinvolto in un piegamento successivo associato a sovrascorimenti fuori sequenza. Nel caso di studio di Tellaro, il sistema di faglie estensionali taglia l'anticlinale di Punta Bianca, essendo associato all'assottigliamento tardo-orogenico del settore interno del prisma Appenninico. Il caso di Studio della Lunigiana include entrambi gli scenari precedenti.

NOTE PER IL LETTORE

Questo manoscritto è organizzato in diversi capitoli, secondo l'ordine seguente.

1) RIASSUNTO ESTESO

Riassunto in lingua italiana dei capitoli successivi.

2) INTRODUCTION

Breve introduzione, per inquadrare il contesto dei prismi orogenici, la problematica delle deformazioni estensionali sin-contrazionali, e la loro importanza.

3) OROGENIC WEDGE EVOLUTIONARY PROCESSES

Descrizione dei prismi orogenici e caratterizzazione, tramite la modellizzazione analogica, dei principali processi che guidano la loro evoluzione a lungo termine.

4) JABAL QUSAYBAH ANTICLINE

Caso di studio del Jabal Qusaybah (Montagne dell'Oman), che permette di investigare le interazioni tra deformazioni contrazionali ed estensionali durante lo sviluppo di una struttura anticlinale, ubicata nella porzione frontale di un prisma orogenico, a livelli strutturali superficiali.

5) VAL DI LIMA FAULT SYSTEM

Caso di studio del sistema di faglie della Val di Lima (Appennino Settentrionale), che permette di studiare strutture a grande scala sviluppate in condizioni non metamorfiche, all'interno di un settore di prisma ciclicamente affetto da deformazioni contrazionali ed estensionali.

6) TELLARO DETACHMENT

Caso di studio del Detachment di Tellaro (Appennino Settentrionale), che permette di studiare una struttura estensionale a basso angolo tardo-orogenica, che caratterizza il settore interno di un prisma.

7) LUNIGIANA FAULT SYSTEM

Caso di studio della fossa tettonica della Lunigiana (Appennino Settentrionale), che permette di studiare le strutture estensionali ad alto angolo tardo-orogeniche che caratterizzano la porzione interna di un prisma di accrezione.

8) GENERAL CONCLUSIONS

Capitolo finale in cui vengono evidenziate le conclusioni generali di questo lavoro.

BIBLIOGRAPHY Lista dei riferimenti bibliografici citati in tutti i precedenti capitoli.

Le attività di ricerca si sono svolte nell'ambito di vari progetti, grazie anche a collaborazioni internazionali con colleghi nel mondo dell'Università e dell'Industria.

Gli esperimenti di modellizzazione analogica sono stati eseguiti presso il laboratorio Géosciences Montpellier dell'Université Montpellier II (Francia), con la supervisione del Dott. Jacques Malavieille.

Il caso di studio del Jabal Qusaybah è stato studiato nell'ambito di una convenzione tra il gruppo "Carbonate Reservoir Appraisal and Production Geology Team" di Shell Global Solutions (Olanda), l'Università degli Studi di Parma e la Katholieke Universiteit Leuven (Belgio). Le attività di ricerca sull'evoluzione dei paleofluidi sono state svolte in collaborazione con il Prof. Rudy Swennen e la Dott.ssa Mahtab Mozafari (KU Leuven).

I casi di studio del sistema di faglie della Val di Lima e del Detachment di Tellaro sono stati studiati nell'ambito del progetto PRIN "La deformazione nelle rocce carbonatiche: implicazioni per la caratterizzazione e la modellizzazione dei serbatoi naturali di geofluidi" finanziato dal Ministero Italiano dell'Educazione, Università e Ricerca (PRIN 2009; Coordinatore Nazionale: Prof. E. Tondi, Università di Camerino; Coordinatore Locale: Prof. F. Storti, Università di Parma). Le attività di ricerca sull'evoluzione dei paleofluidi sono state svolte in collaborazione con il Prof. Philippe Muchez e il Prof. Rudy Swennen (KU Leuven).

Il caso di studio del sistema di faglie della Lunigiana è stato studiato in collaborazione con ENI Exploration & Production, che ha permesso di consultare e interpretare i dati sismici su workstation presso i loro uffici di Milano. Il lavoro di interpretazione sismica è stato eseguito in collaborazione con il Dott. Mirko Carlini e il Dott. Andrea Artoni (Università di Parma).

Il presente elaborato è redatto prevalentemente in lingua inglese, per favorirne la sua circolazione nella comunità scientifica internazionale e per consentire la pubblicazione di alcuni capitoli nella forma di articoli su riviste scientifiche di settore. Solo il primo capitolo è scritto in lingua italiana, e costituisce un breve riassunto dei contenuti successivi.

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Parte I

VERSIONE ITALIANA

RIASSUNTO ESTESO

1.1 INTRODUZIONE GENERALE

Nell'ambito di questa Tesi di Dottorato viene studiata la sovrapposizione di deformazioni estensionali e compressive durante lo sviluppo di prismi orogenici. Per una migliore comprensione di questo argomento, viene inizialmente fornita un'introduzione riguardo i prismi orogenici. Viene brevemente introdotto il contesto geodinamico in cui i prismi si sviluppano, e vengono caratterizzati i processi principali che ne influenzano l'evoluzione. In questo contesto, vengono presentati alcuni esperimenti di modellizzazione analogica che permettono di caratterizzare meglio alcuni processi e la loro influenza reciproca.

Successivamente, l'argomento delle deformazioni estensionali sviluppate in un contesto regionale contrazionale è affrontato tramite l'analisi dettagliata di alcuni casi di studio. Gli esempi selezionati sono ubicati in due diversi prismi orogenici, le Montagne dell'Oman (Sultanato dell'Oman) e l'Appennino Settentrionale (Italia). I casi selezionati includono strutture a diversa scala, sviluppate a differenti livelli strutturali. Inoltre, sono incluse strutture estensionali sia ad alto angolo ("Andersoniane") che strutture a basso angolo ("misorientate").

I casi di studio sono investigati con un approccio multidisciplinare che comprende: interpretazione di linee sismiche; analisi delle strutture a grande scala e investigazione del loro significato nel contesto regionale; realizzazione di carte geologico-strutturali; analisi strutturale alla scala dell'affioramento e a quella microscopica; analisi petrografica, mineralogica, geochimica, e microtermometrica di rocce di faglia e vene sintettoniche.

1.1.1 I prismi orogenici

Nei margini convergenti, il movimento delle placche è bilanciato da un processo chiamato subduzione, in cui una placca (detta *placca inferiore*) scivola al di sotto dell'altra (detta *placca superiore*) e si immerge nell'astenosfera (*subduce*). Durante la subduzione, le interazioni fra le due placche sono caratterizzate da differenti processi tettonici, gravitazionali, e superficiali, che possono essere divisi in due gruppi:

- processi erosivi, che provocano il trasferimento di materiale dalla placca superiore a quella inferiore;
- processi accrezionali, che provocano il trasferimento di materiale dalla placca inferiore a quella superiore.

Quando i processi erosivi sono dominanti si sviluppa un *margin* *convergente erosivo*; al contrario, quando i processi accrezionali sono dominanti, si sviluppa un *margin* *convergente accrezionale* (e.g., Cloos & Shreve, 1988; Von Huene & Lallemand, 1990; Von Huene & Scholl, 1991; Bourgois et al., 1996; Ranero & von Huene R, 2000; Von Huene, 2003; Clift, 2004; Fisher et al., 2004; Vannucchi et al., 2004; Von Huene et al., 2004; Ranero et al., 2008).

La morfologia tipica di un *margin* *convergente accrezionale*, dalla placca inferiore a quella superiore, è caratterizzata da: un avampaese indeformato, una fossa, un prisma orogenico, un bacino sedimentario di avan-arco, un arco vulcanico. Il prisma orogenico, in inglese “*orogenic wedge*”, è un corpo delimitato da faglie ubicato tra la fossa e il basamento della placca superiore. È costituito da materiali raschiati dalla placca inferiore e accreti alla placca superiore (Von Huene & Scholl, 1991). I prismi orogenici sono frequentemente nominati anche “*prismi di accrezione / accretionary wedges*”, essendo il risultato di processi accrezionali, o “*thrust wedges*” essendo dominati da faglie inverse/sovrascorrimenti (*thrust faults*).

La struttura di un tipico prisma di accrezione è ben nota ai geologi da decenni, grazie allo studio di prismi esposti in affioramento, dati di sottosuolo, e modellizzazione numerica o analogica. Il limite esterno del prisma è un sovrascorrimento frontale che separa l'avampaese esterno indeformato dal materiale già accreto alla placca superiore. La faglia del sovrascorrimento frontale ha una geometria “*ramp and flat*” e continua al di sotto del prisma con un segmento a basso angolo. Questo segmento, chiamato scollamento basale, separa il prisma dalla placca in subduzione ed è caratterizzato da deformazione intensa e importante circolazione di fluidi (Moore, 1989).

1.1.1.1 *Processi principali*

I prismi orogenici sono generalmente caratterizzati da processi accrezionali (i.e., *accrezione frontale* e *l'underplating*) tramite cui materiali originariamente appartenenti alla placca inferiore vengono incorporati nella placca superiore in nuove unità tettono-stratigrafiche. Oltre a questo, sono anche attivi processi che dismembrano il prisma, come l'*erosione basale*, l'*erosione superficiale*, e l'*assottigliamento tettonico* (e.g., Cloos & Shreve, 1988; Von Huene & Scholl, 1991; Lallemand et al., 1994; Dominguez et al., 1998b; Clift, 2004; Vannucchi et al., 2004; Von Huene et al., 2004).

L'accrezione frontale determina l'allungamento del prisma tramite l'accrezione di nuovo materiale al fronte. Con la prosecuzione del raccorciamento, infatti, il fronte della deformazione si propaga verso l'avampaese indeformato tramite la nucleazione di una nuova faglia inversa, in posizione più esterna. Ogni nuova faglia diventa temporaneamente il nuovo sovrascorrimento frontale del prisma, mentre le vecchie faglie possono essere disattivate o continuare la loro attività come faglie fuori sequenza (OOS; Morley, 1988). Va sottoli-

neato che l'accrezione frontale non coinvolge tutto il materiale della placca inferiore: una grossa parte del materiale in subduzione non viene accreto al prisma, ma trascinato al di sotto di esso, nel foot-wall dello scollamento principale. Questo processo prende il nome di *underthrusting*.

L'underplating, o accrezione basale, è un processo attivo alla base del prisma, in posizione più interna rispetto all'accrezione frontale (e.g., Elliott & Johnson, 1980; Price, 1981; Platt et al., 1985; Hatcher, 1989; Gutscher et al., 1996; Burkhard & Sommaruga, 1998; Mosar, 1999; Kukowski et al., 2002; Avouac, 2003; Malavieille, 2010; Konstantinovskaya & Malavieille, 2011; Long et al., 2011). Questo processo interessa parte del materiale che ha subito *underthrusting* e lo incorpora alla base del prisma. L'underplating determina lo sviluppo di strutture a duplex e l'aumento dello spessore del prisma. L'aumento di spessore è particolarmente marcato quando il duplex ha una struttura a duomo; questo è il caso, ad esempio, dell'antiformal stacking che causa il sollevamento di un settore di prisma molto localizzato.

L'erosione basale è un processo tettonico attivo alla base del prisma, che comporta il trasferimento di materiale dalla placca superiore a quella inferiore (e.g., Von Huene & Lallemand, 1990; Von Huene & Scholl, 1991; Bourgois et al., 1996; Von Huene, 2003; Clift, 2004). Nei margini convergenti di tipo accrezionale, l'erosione basale generalmente interessa materiali originariamente derivanti dalla placca inferiore, incorporati nel prisma tramite processi accrezionali, e in seguito erosi. Tuttavia, in determinate circostanze (i.e. se il prisma orogenico è piccolo), anche il basamento cristallino della placca superiore può essere eroso (Von Huene & Lallemand, 1990). Come conseguenza dell'erosione basale, tutto il prisma può subire una subsidenza rilevante (e.g., Von Huene & Lallemand, 1990; Clift & MacLeod, 1999; Meschede et al., 1999; Wells, 2003).

L'assottigliamento tettonico implica l'attività di sistemi di faglie che modificano la forma generale del prisma trasferendo materiale da un settore ad un altro (Platt, 1986). Un esempio tipico di questo processo è rappresentato dai sistemi di faglie estensionali a basso angolo che delaminano la porzione superficiale di un prisma orogenico. Tali sistemi di faglie estensionali possono ridurre lo spessore di un settore di prisma trasferendo il materiale in settori adiacenti. L'assottigliamento tettonico, quindi, può non implicare il trasferimento di materiale dalla placca superiore a quella inferiore, a lasciare quindi inalterato il volume complessivo del prisma.

Come conseguenza dei processi accrezionali, lo spessore del prisma aumenta e si sviluppa una morfologia complessa, con un gradiente topografico a grande scala tra il prisma e l'adiacente fossa. Conseguentemente, la porzione superficiale del prisma può essere soggetta a processi erosivi superficiali, che possono essere attivi, con modalità differenti, in ambiente sottomarino o emerso. Il materiale eroso è trasportato per distanze brevi o lunghe, e quindi depositato in ba-

cini al di sopra del prisma o nella fossa. Come risultato dell'azione combinata di erosione superficiale e sedimentazione, quindi, materiali precedentemente accreti al prisma possono essere ritrasferiti alla placca inferiore. In seguito, gli stessi materiali possono essere accreti nuovamente al prisma con l'avanzare del fronte deformativo.

1.1.1.2 *Evoluzione dei prismi orogenici*

I prismi orogenici sono generalmente associati a convergenza su grande scala, ed evolvono quindi in intervalli temporali molto lunghi (e.g., Fossen, 2010). Tutti i processi sopra descritti caratterizzano questa evoluzione a lungo termine, ma la loro attività non è costante nel tempo: si influenzano a vicenda, e sono tutti influenzati dalla presenza di eterogeneità o strutture ereditate nella placca in subduzione e cambiamenti delle condizioni al contorno (i.e., la velocità e l'angolo di convergenza tra le placche, l'angolo di immersione della placca in subduzione, le condizioni climatiche).

Conseguentemente, strutture deformative complesse crescono ed evolvono nel tempo grazie alle interazioni a lungo termine tra processi che modificano le dinamiche dei prismi attraverso il trasferimento di materiale (Elliott, 1976; Chapple, 1978; Storti & McClay, 1995; Mugnier et al., 1997; Konstantinovskaya & Malavieille, 2005; Bonnet et al., 2007; Malavieille, 2010; Perrin et al., 2013). Molti di questi processi, e la loro reciproca influenza, possono essere studiati efficacemente con esperimenti di modellizzazione analogica. In particolare, nel capitolo 3, viene studiata l'evoluzione di prismi interessati da accrezione frontale, underplating e erosione superficiale. La sedimentazione sin-tettonica e l'erosione basale non sono state prese in considerazione in questi esperimenti in quanto già ampiamente trattati in ricerche precedenti (sedimentazione: e.g., Cobbold et al., 1989, 1993; Malavieille et al., 1993; Baby et al., 1995; Storti & McClay, 1995; Persson & Sokoutis, 2002; Bigi et al., 2010; erosione basale: e.g., Lallemand et al., 1992, 1994; Malavieille et al., 1992; Mandal et al., 1997; Dominguez et al., 1998a,b, 2000; Nieuwland et al., 2000; Marques & Cobbold, 2002; Hampel et al., 2004; Gutscher et al., 2009). L'assottigliamento tettonico non è stato investigato per via di limiti intrinseci della metodologia sperimentale.

Dal punto di vista di tutto il prisma, le interazioni a grande scala tra processi accrezionali e erosivi sono vincolati nel contesto meccanico della teoria del *critical taper*, che permette di stabilire una relazione tra le caratteristiche meccaniche e geometriche dei prismi orogenici (Davis et al., 1983; Dahlen et al., 1984; Dahlen, 1984, 1990). Secondo questa teoria, in particolare, i prismi tenderebbero a raggiungere e mantenere una pendenza superficiale meccanicamente stabile, determinata dalla resistenza dei materiali che costituiscono il prisma stesso, dalla pendenza dello scollamento basale, e dal coefficiente di frizione lungo di esso. Il limite maggiore della teoria del *critical taper* è che essa si limita a descrivere la geometria a grande scala del pri-

sma, senza dare informazioni riguardo come questa geometria venga raggiunta: non tiene conto dei processi attivi all'interno del prisma, che permettono di raggiungere il profilo stabile. Nonostante questo, l'applicazione del critical taper ai prismi orogenici e alle catene *fold and thrust* è stato uno dei più significativi progressi recenti nello studio della loro meccanica a grande scala (e.g., Davis et al., 1983; Dahlen et al., 1984; Dahlen, 1984, 1990; Lallemand et al., 1994).

Applicando la teoria del critical taper all'evoluzione a lungo termine dei prismi orogenici, è possibile interpretare le interazioni tra i processi accrezionali/sedimentari e quelli estensionali/erosivi in un contesto meccanico coerente. I processi accrezionali e sedimentari, aumentando lo spessore del prisma, possono ristabilire un profilo critico quando/dove il prisma, o una singola sua porzione, sia in condizioni sottocritiche. I processi erosivi ed estensionali, riducendo lo spessore del prisma, possono ristabilire un profilo critico quando/dove il prisma, o una sua porzione, sia in condizioni supercritiche. Conseguentemente, l'evoluzione a lungo termine dei prismi orogenici è caratterizzata dalla coesistenza di deformazioni contrazionali ed estensionali (e.g., Platt, 1986; Burchfiel et al., 1992; Platt, 1993; Wallis et al., 1993; Crespi et al., 1996; Jones & Strachan, 2000; Andresen et al., 2007).

Da lungo tempo è riconosciuto che le deformazioni estensionali caratterizzano gli stadi finali dell'evoluzione dei prismi orogenici, quando gli sforzi compressivi si attenuano o migrano verso settori più esterni (Lister et al., 1984; Malinverno & Ryan, 1986; Dewey et al., 1989; Carmignani & Kligfield, 1990; Patacca et al., 1990; Doglioni, 1991; Burchfiel et al., 1992; Molnar et al., 1993; Carmignani et al., 1994; Crespi et al., 1996; Jolivet et al., 1996, 1998; Reddy et al., 1999; Brogi, 2004, 2008), eventualmente associati a processi come l'arretramento flessurale della placca in subduzione e l'apertura di bacini di retroarco (Jolivet et al., 1991, 1994, 1998; Faccenna et al., 1997; Gueguen et al., 1998; Faccenna et al., 2001; Brun & Faccenna, 2008). Deformazioni compressive ed estensionali, tuttavia, possono anche coesistere in stadi evolutivi precedenti, associati a: i) la crescita di una singola piega, e indotte dalla sua evoluzione cinematica; ii) la storia orogenica dell'intero prisma, attivate per ristabilire un profilo meccanicamente stabile qualora il prisma sia in condizione supercritiche (Davis et al., 1983; Platt, 1986), come predetto dalla teoria del profilo critico.

1.1.2 Deformazioni estensionali sin-contrazionali

Come spiegato nella sezione precedente, l'occorrenza di eventi estensionali durante l'evoluzione di prismi orogenici è frequente. Le conseguenti strutture deformative estensionali sin-contrazionali possono essere associate a processi di rilevanza scientifica ed economica, meritando quindi di essere investigate nel dettaglio.

Ad una scala locale, strutture compressive ed estensionali a scala

differente posso interagire durante lo sviluppo di una piega associata ad un sovrascorrimento, come nel caso dell'anticlinale di Jabal Qusaybah (capitolo 4). Nel Jabal Qusaybah, infatti, una piega anticlinale chilometrica è affetta da faglie estensionali metriche - ettometriche che: i) accomodano l'estensione di estradosso longitudinale e trasversale associata al piegamento; ii) si sviluppano in array transtensivi alla terminazione di faglie trascorrenti.

A scala più grande, stadi importanti di estensione possono risultare da condizioni supercritiche in un settore specifico di un prisma orogenico, mentre settori adiacenti sono caratterizzati da condizioni sottocritiche e deformazioni compressive. Questo è il caso della Val di Lima (capitolo 5), in cui deformazioni estensionali interessano la porzione superficiale del prisma Appenninico, e sono attivate da un sollevamento localizzato di una porzione di prisma. In più, la Val di Lima evidenzia la complessità strutturale dovuta alla sovrapposizione multipla di deformazioni compressive ed estensionali che interessano ciclicamente la stessa area. Il Detachment di Tellaro (capitolo 6) è ubicato invece in una porzione più interna dello stesso prisma orogenico, in un settore in cui le deformazioni estensionali sono più sviluppate, e includono: i) deformazioni estensionali sin-contrazionali, che hanno portato ad un marcato assottigliamento della porzione superficiale del prisma e contribuito all'esumazione di unità metamorfiche; ii) sistemi di faglie estensionali tardo- e post-orogeniche, associate alla migrazione del fronte deformativo e all'apertura del bacino Tirrenico.

Infine, il caso di studio della Lunigiana (capitolo 7) investiga le interazioni tra deformazioni estensionali e contrazionali a scala regionale. Questo caso di studio, in cui sono presenti strutture estensionali sin-orogeniche e post-orogeniche, fornisce un contesto regionale in cui è possibile includere i casi di studio della Val di Lima e del Detachment di Tellaro.

Concludendo, va sottolineato che le zone di faglia estensionali influenzano la circolazione dei fluidi nella crosta superiore e modificano le proprietà meccaniche dei prismi orogenici favorendo la sovrappressione dei fluidi in profondità e lo sviluppo di minerali idrati a basso coefficiente di frizione (Hubbert & Rubey, 1959; Davis et al., 1983; Dahlen et al., 1984; Sibson et al., 1988; Dahlen, 1990; Moore & Vrolijk, 1992; Kirschner & Kennedy, 2001; Collettini & Holdsworth, 2004; Fagereng et al., 2010; Lacroix et al., 2013). L'architettura delle zone di faglia, infatti, comprende una rete idraulica complessa di barriere, condotti, e sistemi misti barriera-condotto (Caine et al., 1996; Storti et al., 2003) che può: i) costituire una via preferenziale di migrazione dei fluidi, permettendo l'infiltrazione in profondità di fluidi di origine superficiale (Wickham & Taylor, 1987; Fricke et al., 1992; Morrison, 1994; Morrison & Anderson, 1998; Bebout et al., 2001; Verhaert et al., 2004, 2009; Mulch et al., 2004, 2006; Person et al., 2007; Ingebritsen & Manning, 2010; Gébelin et al., 2011; Gottardi et al., 2011); ii) costituire barriere impermeabili che favoriscono la compartimentalizzazione

dei fluidi in profondità e innescano condizioni di sovrappressione (Sibson, 1981; Parry & Bruhn, 1990; Sibson, 1992; Gudmundsson, 1999; Sibson, 2000; Gudmundsson, 2001; Collettini & Barchi, 2002; Collettini et al., 2006a; Cobbold et al., 2009). In queste circostanze, le zone di faglia estensionali possono favorire mineralizzazioni di importanza economica rilevante.

Inoltre, la contemporanea attività di deformazioni estensionali e compressionali, rispettivamente nelle porzioni superficiali e profonde di un prisma orogenico, può favorire fenomeni geologici complessi come l'esumazione sin-contrazionale di unità metamorfiche (Lister et al., 1984; Platt, 1993; Crespi et al., 1996; Jolivet et al., 1996, 1998, 2003; Brun & Faccenna, 2008).

1.2 PROCESSI EVOLUTIVI DEI PRISMI OROGENICI

Questa sezione riassume il capitolo 3, costituito dalla pubblicazione "Impact of erosion and décollements on large-scale faulting and folding in orogenic wedges: analogue models and case studies" pubblicato da PERRIN C., CLEMENZI L., MALAVIEILLE J., MOLLI G., TABOADA A., AND DOMINGUEZ S. sulla rivista *Journal of the Geological Society*.

La modellizzazione analogica ha permesso di evidenziare i processi di che caratterizzano l'evoluzione a lungo termine dei prismi orogenici. Sebbene i limiti intrinseci del metodo sperimentale non permettano di investigare direttamente le deformazioni estensionali sin-contrazionali, questo capitolo riveste comunque un ruolo importante nell'ambito della Tesi di Dottorato. Viene infatti delineato il contesto geodinamico in cui sono inquadrati i casi di studio specifici investigati nei capitoli successivi, e vengono caratterizzati i rapporti tra i processi accrezionali e quelli erosivi.

1.2.1 Risultati sperimentali

I meccanismi di deformazione, la cinematica a lungo termine e l'evoluzione di catene tipo "fold and thrust" soggette a erosione superficiale sono studiati tramite esperimenti analogici 2D a grande convergenza. I parametri di prim'ordine testati negli esperimenti includono (i) livelli di scollamento e livelli plastici inseriti in diverse posizioni nel materiale analogico e (ii) erosione superficiale contemporanea alla convergenza.

I livelli di scollamento, a seconda delle loro posizione, possono favorire la partizione della deformazione e la conseguente contemporanea attivazione di due differenti processi tettonici: i) l'accrezione basale ("underplating"), attiva nella porzione interna del prisma, che determina l'ispessimento del prisma tramite l'impilamento di diverse unità tettoniche alla sua base; (ii) l'accrezione frontale, che determina un aumento della lunghezza del prisma tramite la propagazione del fronte di deformazione verso l'avampaese. Lo sviluppo di strutture a

duplex di tipo “antiformal stack” è controllato dall’accrezione basale e mostra una ciclicità a piccola e grande scala.

La presenza di sottili livelli plastici induce processi di piegamento che sono studiati alla scala di tutto il prisma. Pieghe ribaltate e rovesciate, con lunghi fianchi inversi, si sviluppano in un regime di deformazione asimmetrico indotto dal taglio, tramite il progressivo srotolamento della cerniera sinclinale. L’erosione superficiale e l’accrezione basale inducono un’ulteriore rotazione passiva dei fianchi delle pieghe, fino alla loro orizzontalizzazione.

In accordo con le teorie del “critical taper model” si è osservato che i prismi di accrezione tendono a mantenere un profilo critico determinato dal comportamento meccanico dei materiali che lo costituiscono, dalla pendenza del livello di scollamento basale e dal coefficiente di attrito lungo di essa (Davis et al., 1983; Platt, 1986). In assenza di processi superficiali, il profilo critico viene mantenuto dalla contemporanea attività di strutture che determinano un ispessimento del prisma (accrezione basale e sovrascorrimenti fuori-sequenza) e strutture che determinano un aumento della sua lunghezza (accrezione frontale). Tuttavia, quando i processi superficiali sono attivi, le interazioni tra la tettonica e i processi superficiali influenzano questo comportamento. In particolare, l’erosione superficiale nella porzione interna del prisma contribuisce a mantenere il prisma in condizione sotto-critiche e favorisce i processi tettonici di accrezione basale.

I risultati di questa ricerca contribuiscono a caratterizzare le interazioni tra processi accrezionali ed erosivi, e la loro influenza reciproca nel contesto dell’evoluzione a lungo termine dei prismi orogenici. L’interpretazione dei risultati sperimentali fornisce inoltre spunti di riflessione che permettono di discutere i meccanismi responsabili della formazione di strutture a grande scala (i.e., impilamento di unità tipo “antiformal stack”, “klippen” e pieghe rovesciate chilometriche) che sono osservate in diverse catene orogeniche tipo la Montagna Nera (Francia), la catena Varisica Galiziana (Spagna) e l’Appennino Settentrionale (Italia).

1.3 ANTICLINALE DI JABAL QUSAYBAH

Questa sessione riassume il manoscritto “Complex fault-fold interactions during the growth of the Jabal Qusaybah inversion anticline at western tip of the Salakh Arch, Oman.”, autori: STORTI F., BALSAMO F., CLEMENZI L., MOZAFARI M., AL-KINDY M.H.N., SOLUM J., SWENNEN R., TABERNER C., AND TUECKMANTEL C., in corso di sottomissione alla rivista *Tectonics*.

Questo caso di studio ha permesso di analizzare le relazioni tra strutture contrazionali ed estensionali alla scala di una singola struttura plicativa, al fronte di un prisma di accrezione. Sono presenti strutture caratterizzate da riattivazioni multiple, in compressione ed estensione, e strutture estensionali formatesi come conseguenza locale di

strutture compressive a scala maggiore.

1.3.1 *Introduzione*

Il Jabal Qusaybah è un rilievo morfologico di dimensioni $\sim 10 \times 3$ km, allungato in direzione W-E e sopraelevato di circa 300 rispetto alla pianura circostante. Costituisce la terminazione occidentale del Salakh Arc, un allineamento di rilievi associati a pieghe anticlinali chilometriche che portano in affioramento la successione carbonatica delle formazioni Natih, del Cretaceo Inferiore. Lo sviluppo di tali anticlinali è associato all'attività del sovrascorrimento frontale del prisma di accrezione delle Montagne dell'Oman. Il rilievo di Jabal Qusaybah, in particolare, è associato a una piega anticlinale con asse orientato in direzione W-E e caratterizzata dalla presenza di un rigonfiamento nella porzione centrale, in corrispondenza della culminazione assiale. L'interpretazione di un profilo sismico commerciale ha evidenziato come il sovrascorrimento responsabile dell'anticlinale derivi dalla riattivazione positiva di una precedente faglia estensionale immergente verso N, con evidenze di attività estensionale sia Paleozoica che Cretacica. L'intera struttura anticlinale è affetta da numerose faglie, a scala da metrica a chilometrica, appartenenti a sistemi con diversa orientazione e cinematica, che sono stati investigati nel dettaglio con un approccio multidisciplinare.

1.3.2 *Risultati*

Il rilievo geologico-strutturale sul terreno ha permesso di individuare e mappare numerosi sistemi di faglie che interessano la struttura anticlinale. Sono presenti: i) faglie trascorrenti sinistre con orientazione $\sim$ NE-SW, ii) faglie trascorrenti destre con orientazione $\sim$ NW-SE; iii) faglie ad orientazione W-E che mostrano evidenze di riattivazioni multiple con cinematica sia estensionale che contrazionale e talvolta attività estensionale sinsedimentaria; iv) faglie normali con orientazione $\sim$ N-S, ubicate in prevalenza nella zona del rigonfiamento centrale e caratterizzate da rapporti rigetto/lunghezza molto elevati. La maggior parte delle faglie presentano nuclei dilatati riempiti di calcite; la dilatazione è particolarmente rilevante nella zona del rigonfiamento centrale, dove i nuclei dilatati hanno dimensioni plurimetriche.

Le vene sintettoniche e i nuclei delle faglie sono riempiti da numerose generazioni di calcite, che sono state suddivise in tre gruppi in base alle loro caratteristiche petrografiche, geochimiche, e ai rapporti di intersezione; ogni gruppo di riempimenti di calcite si trova prevalentemente associato a determinate strutture.

1.3.3 *Discussione*

L'analisi multidisciplinare ha permesso di suddividere l'evoluzione dell'anticlinale di Jabal Qusaybah e delle strutture associate in vari stadi successivi. Le prime fase evolutive, durante la sedimentazione delle formazioni Natih e le prime fasi della diagenesi, sono caratterizzate dallo sviluppo di faglie estensionali sinsedimentarie ad orientazione E-W, vene sintettoniche associate, e strutture da compattazione quali stiloliti parallele alla stratificazione e vene ortogonali. In seguito, l'instaurarsi di un campo di stress regionale contrazionale più intenso determina l'attivazione delle faglie trascorrenti sinistre NE-SW e, conseguentemente, si ha la formazione di faglie estensionali orientate circa N-S (horse-tail trasversali alla terminazione delle faglie trascorrenti) e la riattivazione in compressione di alcune faglie E-W. Contemporaneamente, altre faglie E-W normali sono ancora attive, testimoniando l'estensione di estradosso associata ad un piegamento embrionale. Successivamente, l'amplificarsi del piegamento e accentua l'estensione di estradosso, che innesca una fase di dilatanza generalizzata in cui: i) le faglie E-W e quelle N-S sono attive in estensione; ii) le faglie NE-SW sono attive in trascorrenza sinistra; iii) i nuclei di tutte le faglie sono dilatati e riempiti di calcite. In seguito, col proseguire della deformazione, i riempimenti di calcite depositati nei nuclei delle faglie vengono rideformati e brecciati, mentre alcune faglie E-W sono invertite positivamente. Infine, l'ultima fase di evoluzione della struttura è caratterizzata dallo sviluppo del rigonfiamento nella porzione centrale della piega. La conseguente estensione longitudinale determina, in questo settore, lo sviluppo delle faglie estensionali N-S di dimensioni maggiori e la dilatazione dei nuclei delle faglie fino a dimensioni metriche. Questo quadro evolutivo testimonia la complessità delle relazioni tra compressione, amplificazione del piegamento ed estensione di estradosso, che ciclicamente porta a riattivazioni in compressione o estensione delle principali strutture.

1.3.4 *Conclusioni*

L'analisi strutturale, petrografica e geochimica delle rocce di faglia (riempimenti dei nuclei delle faglie e vene sintettoniche) ha dimostrato che il piegamento e l'attività delle faglie estensionali e trascorrenti sono avvenuti contemporaneamente durante lo sviluppo dell'anticlinale di Jabal Qusaybah. Ha inoltre permesso di caratterizzare le condizioni P-T della deformazione, l'origine dei fluidi coinvolti e determinare la cronologia relativa dell'attività delle faglie. In particolare, il modello evolutivo proposto evidenzia la complessità delle strutture che si possono sviluppare grazie alle interazioni tra deformazioni compressive ed estensionali. Il pattern deformativo di Jabal Qusaybah è più complesso di quello predetto dai modelli con cui vengono tradizionalmente descritte le strutture anticlinali di avampaese e la frattura-

razione associata, sottolineando quindi l'importanza dello studio di analoghi di terreno per applicazioni industriali quali lo sfruttamento di idrocarburi.

1.4 SISTEMA DI FAGLIE DELLA VAL DI LIMA

Questa sezione riassume il capitolo 5, costituito dal manoscritto "Extensional deformation structures within a convergent orogen: The Val di Lima low-angle normal fault system (Northern Apennines, Italy)", sottomesso da CLEMENZI L., MOLLI G., STORTI F., MUCHEZ P., SWENNEN R., TORELLI L. alla rivista *Journal of Structural Geology*.

Questo caso di studio ha permesso di investigare le interazioni tra strutture contrazionali ed estensionali a scala plurichilometrica, nella porzione centrale di un prisma di accrezione. È stato studiato, nel dettaglio, un sistema di faglie estensionali a basso angolo sviluppato in un contesto regionale contrazionale e successivamente deformato da strutture compressive.

1.4.1 Introduzione

L'area di studio è ubicata nella porzione centrale del prisma di accrezione dell'Appennino Settentrionale, in prossimità del crinale Tosco-Emiliano e si colloca in posizione esterna rispetto alla zona di affioramento del nucleo metamorfico Apuano. Quest'area è caratterizzata dalla completa esposizione di tutte le formazioni della successione Toscana non metamorfica, che comprende: le brecce tettoniche basali del Calcare Cavernoso, la successione sedimentaria Triassico-Eocenica, spessa circa 2 km e prevalentemente carbonatica, i depositi siliciclastici di avanfossa Chattiani - Aquitaniani del Macigno. Dal punto di vista strutturale, la successione Toscana è localmente coinvolta in una piega chilometrica isoclinale ribaltata, con il piano assiale orizzontale localizzato nel Calcare Cavernoso. Il fianco normale della piega è delaminato da un sistema di faglie a basso angolo che determina, localmente, l'elisione completa di alcune formazioni. L'intera struttura è quindi deformata da pieghe aperte NE-vergenti, che deformano i fianchi della piega isoclinale e il sistema di faglie a basso angolo. Infine, l'area di studio è interessata da faglie tardive ad alto angolo, con cinematica estensionale e trascorrente.

1.4.2 Risultati

La struttura a grande scala della Val di Lima è stata vincolata tramite l'analisi di un profilo sismico commerciale. L'interpretazione del dato di sottosuolo ha evidenziato che: i) la piega isoclinale coinvolge solo le formazioni Triassico-Eoceniche della successione Toscana, mentre il flysh del Macigno non è presente nel fianco inverso; ii) la struttura-

zione a pieghe aperte è associata all'attività di un sovrascorrimento tardivo, NE-vergente, che taglia i fianchi della piega isoclinale.

Gli assetti geometrici del sistema di faglie a basso angolo sono stati vincolati con un rilievo geologico-strutturale sul terreno, mentre la sua configurazione originaria è stata ricostruita retrodeformando le strutture sovrainposte. Il sistema di faglie ha un'architettura "a scalinata" con segmenti principali a basso angolo collegati da splay secondari con geometrie anastomizzate. La direzione di trasporto tettonico è verso NE, con locali variazioni nell'intervallo N020 – N120. Il segmento di faglia a basso angolo principale, generalmente, determina l'elisione completa della Scaglia Toscana e porta il Macigno a contatto diretto con la Maiolica. Architetture più complesse sono comunque presenti e includono la presenza di due o più segmenti di faglia a basso angolo giustapposti, che delimitano lenti tettoniche di spessore variabile (10 – 100 m). Localmente l'elisione tettonica può essere più marcata (es: contatto diretto tra il Macigno e le formazioni più profonde della successione Toscana), oppure elisioni e ripetizioni tettoniche possono coesistere nella stessa area.

La zona di danneggiamento associata alle faglie a basso angolo presenta importanti variazioni che dipendono non solo dalle diverse litologie coinvolte, ma anche dalla posizione strutturale (hangingwall o footwall). In particolare, l'architettura della zona di danneggiamento è stata caratterizzata in due settori differenti. Nel sito 1 i carbonati della formazione Maiolica, all'hangingwall, mostrano una zona di danneggiamento metrica con una tessitura brecciata e sono interessati da vene di calcite parallele o ortogonali al trasporto tettonico. Le argilliti della Scaglia Toscana, al footwall, mostrano una zona di danneggiamento delocalizzata, spesso alcune decine di metri, caratterizzata da una foliazione parallela alla faglia e quindi a basso angolo rispetto alla stratificazione. Il nucleo della faglia è costituito da un gouge sviluppato prevalentemente a spese della Scaglia, ma che include anche noduli di selce provenienti dalla Maiolica; tali noduli sono interessati da vene di quarzo orientate ortogonali alla direzione di trasporto tettonico. Nel sito 2 le arenarie del Macigno, all'hangingwall, sono caratterizzate da una zona di danneggiamento metrica, che comprende: i) vicino al contatto tettonico principale, alternanze centimetriche di cataclasi foliate, brecce e roccia relativamente indeformata; ii) a distanza crescente dal contatto tettonico principale, domini di roccia relativamente indeformata e zone di taglio millimetriche molto spaziate tra loro. Il nucleo della faglia è spesso pochi centimetri e costituito da una cataclasi foliata, caratterizzata da una tessitura matrice-sostenuta ricca di ossidi e fillosilicati, con microstrutture (P- R- e S-shears) consistenti con la direzione di trasporto tettonico. I carbonati della Maiolica, al footwall, sono caratterizzati da vene sintettoniche la cui abbondanza aumenta avvicinandosi al contatto tettonico principale, fino a raggiungere valori estremamente elevati. In prossimità della superficie di scorrimento è presente una struttura deformativa

particolare: i) la roccia è costituita quasi interamente da vene sintettoniche di calcite mentre l'host rock è dolomitizzato e preservato solo in piccoli frammenti; ii) le vene hanno un'orientazione più a basso angolo rispetto al piano di faglia e sono deformate in un pattern S-C.

Le condizioni di pressione e temperatura del sistema di faglie, e l'evoluzione dei paleofluidi coinvolti nella deformazione, sono stati vincolati studiando le rocce di faglia e le vene sintettoniche riempite di calcite o quarzo. Nella cataclasite foliata al nucleo della faglia (sito 2) l'analisi mineralogica delle argille mostra una cristallinità dell'Illite minore rispetto al Macigno indeformato, indicando che la deformazione è successiva al picco termico registrato dall'arenaria. La microtermometria delle inclusioni fluide nelle vene sintettoniche di quarzo ha indicato temperature di omogeneizzazione nell'intervallo da 150 a 160°C e temperature di fusione del ghiaccio nell'intervallo da -3 a -1°C. La microtermometria delle inclusioni fluide nelle vene sintettoniche di calcite, invece, ha indicato temperature di omogeneizzazione più basse, nell'intervallo da 120 a 130°C e temperature di fusione del ghiaccio nell'intervallo da -5 a -0.5°C; la calcite tuttavia è affetta da abbondante ricristallizzazione. L'analisi degli isotopi stabili (carbonio e ossigeno) è stata effettuata su host rock e vene di calcite in campioni di Maiolica provenienti dalle zone di faglia (siti 1 e 2) e da una zona relativamente indeformata (sito 3). I risultati analitici mostrano valori molto diversi tra gli host rock nei vari settori, e minore variabilità tra le vene di calcite. La Maiolica dolomitizzata del sito 2 mostra valori $\delta^{13}\text{C}$ leggermente negativi (tra -1 e 0‰ V-PDB) e valori $\delta^{18}\text{O}$ molto negativi (tra -7.5 e -6‰ V-PDB); la Maiolica del sito 1 mostra valori $\delta^{13}\text{C}$ positivi (~ +2‰ V-PDB) e valori $\delta^{18}\text{O}$ negativi (~ -5‰ V-PDB); la maiolica del sito 3 mostra valori $\delta^{13}\text{C}$ intermedi (~ +1‰ V-PDB) e valori $\delta^{18}\text{O}$ solo leggermente negativi (~ -1.8‰ V-PDB). Le vene di calcite mostrano generalmente valori $\delta^{13}\text{C}$ leggermente positivi (tra 0.5 e 2‰ V-PDB) e valori $\delta^{18}\text{O}$ variabili da negativi a molto negativi (tra -6 e -3‰ V-PDB); solo alcune vene si discostano da questi valori, e mostrano composizione simile al relativo host rock. La microscopia ottica in catodoluminescenza mostra evidenze di ricristallizzazione delle vene di calcite e della Maiolica dolomitizzata.

1.4.3 *Discussione*

Il picco termico registrato dalla successione Toscana è associato al suo massimo seppellimento tettonico, stimato a circa 250°C e 7 km di profondità, con un gradiente geotermico di 30°C/km (Molli et al., 2011, e riferimenti inclusi). La cristallinità dell'Illite nella cataclasite foliata indica che il sistema di faglie a basso angolo è stato attivo in un momento successivo, durante l'esumazione della successione Toscana. Considerando il gradiente geotermico di 30°C/km e le analisi microtermometriche nelle vene di quarzo sintettoniche, si può: i) vincolare l'attività della faglia a circa 180°C e 5 km di profondità; ii) ipotizzare

il coinvolgimento di fluidi di origine superficiale a bassa salinità, infiltrati in profondità nella crosta grazie alla permeabilità associata al sistema di faglie. L'elevata permeabilità ha permesso anche un'abbondante circolazione di fluidi in fasi tettoniche successive, favorendo la ricristallizzazione di host rock e vene di calcite a condizioni P-T decrescenti (i.e., $\sim 150^{\circ}\text{C}$ e 3 km di profondità per le vene di calcite dei siti 1 e 2). La bassa salinità e l'arricchimento isotopico ($\delta^{18}\text{O}$ da +12 a +14‰ SMOW) del fluido associato alla ricristallizzazione sono compatibili con un'acqua di origine superficiale che abbia interagito abbondantemente con rocce siliciclastiche durante l'infiltrazione nella crosta.

L'integrazione dei risultati esposti ha permesso di proporre un modello concettuale per spiegare l'evoluzione strutturale della Val di Lima nel contesto regionale del prisma Appenninico. Il piegamento iniziale della successione Toscana sarebbe legato ad una prima fase di sovrascorrimento mentre una successiva amplificazione del piegamento avrebbe permesso lo sviluppo di una geometria isoclinale con fianco inverso chilometrico. Successivamente, processi di accrezione basale (duplexing e antiformal stacking) avrebbero causato il sollevamento localizzato della porzione più interna del prisma, con due conseguenze: i) una rotazione passiva, fino all'orizzontalizzazione, della piega isoclinale; ii) la modifica del profilo del prisma fino al raggiungimento di condizioni supercritiche, meccanicamente instabili. Tale instabilità meccanica avrebbe innescato deformazioni estensionali locali, tra cui il sistema di faglie della Val di Lima, che avrebbero assottigliato le porzioni superficiali del prisma e contribuito a ristabilire un profilo meccanicamente stabile. In un momento successivo, il settore della Val di Lima è nuovamente interessato da deformazioni contrazionali legate all'attivazione di un sovrascorrimento di origine profonda; la deformazione di hangingwall associata causa lo sviluppo delle pieghe aperte con asimmetria a "Z" e l'antiforme frontale. Infine, le faglie estensionali ad alto angolo si sviluppano negli ultimi momenti della storia evolutiva del prisma di accrezione, e tagliano tutte le strutture precedenti. Nella fase l'architettura della zona di danneggiamento del sistema di faglie della Val di Lima, a comportamento misto condotto-barriera, avrebbe modificato la circolazione dei fluidi nella crosta superiore, determinando: i) l'infiltrazione di fluidi di origine superficiale fino a ~ 5 km di profondità; ii) locali condizioni di sovrappressione idraulica, capaci di influenzare i processi di deformazione attivi, favorire la localizzazione della deformazione e lo sviluppo di particolari strutture deformative (i.e., vene S-C nella Maiolica al footwall del sito 2).

Una datazione diretta del sistema di faglie estensionali a basso angolo non è disponibile, ma l'attività delle faglie può essere vincolata nel tempo grazie al contesto geologico regionale. Infatti, le fasi compressive responsabili del piegamento iniziale coinvolgono i depositi di avanfossa del basso Miocenici del Macigno e dell'Unità Cervaro-

la (Chicchi & Plesi, 1991, 1992; Plesi et al., 2002). Le faglie ad alto angolo tardo orogeniche sono associate all'apertura del bacino della Garfagnana, in cui è presente un record sedimentario a partire dal basso Pliocene (Bartolini & Bortolotti, 1971; Calistri, 1974; Nardi et al., 1986, 1987; Dallan et al., 1981; Fazzuoli et al., 1998). Conseguentemente, l'attività del sistema di faglie estensionali a basso angolo e il suo piegamento dovuto a sovrascorrimenti "fuori-sequenza" possono essere ragionevolmente vincolati al tardo Miocene.

1.4.4 Conclusioni

Il caso di studio della Val di Lima è caratterizzato dalla sovrapposizione reciproca di strutture deformative contrazionali ed estensionali sviluppate durante l'evoluzione a lungo termine del prisma di accrezione Appenninico. Il sistema di faglie estensionali a basso angolo, in particolare, indica l'occorrenza di un pulso estensionale durante l'evoluzione contrazione del prisma. Analoghi pulsanti estensionali possono essere comuni durante l'evoluzione di molti prismi di accrezione e le strutture deformative associate possono modificare in modo sostanziale la circolazione dei fluidi nella crosta superiore, fino ad influenzarne il comportamento meccanico.

1.5 IL DETACHMENT DI TELLARO

Questa sezione riassume il manoscritto "Structural anatomy and paleofluid evolution of the Tellaro detachment fault system, Northern Apennines, Italy", sottomesso da CLEMENZI L., STORTI F., BALSAMO F., MOLLI G., ELLAM R., MUCHEZ P., SWENNEN R. alla rivista *Geological Society of America Bulletin*.

Questo caso di studio ha permesso di investigare un sistema di faglie estensionali a basso angolo a grande scala, associato alle deformazioni tardo-orogeniche che interessano il settore interno di un prisma di accrezione. La buona esposizione del Detachment di Tellaro e la qualità degli affioramenti hanno permesso di caratterizzare nel dettaglio l'architettura strutturale del sistema di faglie e la sua evoluzione paleoidrologica.

1.5.1 Introduzione

La porzione interna del prisma di accrezione dell'Appennino settentrionale è assottigliata da faglie estensionali a basso angolo del tardo Miocene e ad alto angolo del Plio-Pleistocene. Tra le prime, il Detachment di Tellaro è particolarmente ben esposto lungo la costa occidentale del promontorio di Punta Bianca. È un sistema di faglie caratterizzato da un trasporto tettonico verso NE, che interessa principalmente le formazioni non metamorfiche della successione Toscana,

riducendone fortemente lo spessore. Localmente, grazie all'attività combinata di faglie estensionali a basso angolo e faglie subsidiarie ad alto angolo, l'originario spessore stratigrafico chilometrico è ridotto di oltre 700 m.

1.5.2 *Analisi strutturale*

Nel sistema di faglie del Detachment di Tellaro, la maggior parte del rigetto è accomodato da segmenti di faglia a basso angolo, che hanno geometrie anastomizzate, lunghezze in traccia ettometriche, e sono sub-orizzontali (pendenza media $\sim 10^\circ$ verso NE). Le faglie subsidiarie ad alto angolo accomodano rigetti minori e si raccordano verso il basso sui segmenti a basso angolo; hanno dimensioni da metriche a decametriche e pendenza variabile a seconda dell'orientazione (valori medi: 70° verso SW e 35° verso NE). L'orientazione attuale del sistema di faglie può essere interessata da rotazione passiva durante gli ultimi stadi di evoluzione e le deformazioni Plio-Pleistoceniche, ma, restorando le faglie subsidiarie ad un'orientazione "Andersoniana", è possibile stimare che l'entità di tale rotazione non superi i 20° . Conseguentemente, i segmenti di faglia principali a basso angolo avrebbero avuto una pendenza originale media di circa 25° verso NE.

L'analisi strutturale di dettaglio ha evidenziato che sono presenti più segmenti di faglia a basso angolo giustapposti, con una spaziatura metrica-decmetrica e il rigetto stratigrafico maggiore generalmente associato al segmento più elevato. La zona di faglia, inoltre, è stata interessata da abbondante circolazione di fluidi, con lo sviluppo di corpi dolomitizzati, pressione-dissoluzione e formazione di abbondanti vene sintettoniche riempite di calcite o dolomite. I corpi dolomitizzati hanno una forma tabulare (estensione laterale decmetrica e spessore metrico), sono discordanti rispetto alla stratificazione delle formazioni circostanti, e si trovano associati ai segmenti di faglia a basso angolo maggiori; sono brecciati e tagliati dalle faglie subsidiarie ad alto angolo. Le vene di dolomite sono presenti solo negli host rock dolomitici, subverticali, e orientate preferenzialmente ortogonali al trasporto tettonico. Le vene di calcite sono presenti in tutte le zone di faglia, con orientazioni variabili e relazioni di intersezione non sistematiche. Inoltre, la zona di danneggiamento di footwall vicino ai segmenti di faglia a basso angolo principali è caratterizzata da vene molto corte, con forma fortemente irregolare.

1.5.3 *Analisi petrografica, geochimica e delle microstrutture*

L'analisi petrografica/microstrutturale delle vene sintettoniche ha rilevato la presenza, nei riempimenti, di una generazione di dolomite e diverse generazioni di calcite; queste ultime sono state classificate in due gruppi in base ai rapporti di intersezione e alla presenza di clivaggio nei cristalli. Le vene del "gruppo 1" sono caratterizzate da

cristalli con twinning di tipo I, e sono sistematicamente tagliate da vene del "gruppo 2" con cristalli limpidi, senza twinning. L'analisi in catodoluminescenza ha permesso di individuare, all'interno di ognuno dei due gruppi, multiple generazioni di riempimento, non associate a una direzione preferenziale. Generalmente ogni singola vena è caratterizzata da una sola generazione di riempimento; esempi di riaperture multiple, con generazioni successive, sono più rari.

L'analisi degli isotopi stabili (ossigeno e carbonio) è stata effettuata mediante microcampionamento di host rock, vene di calcite, vene di dolomite, e brecce dolomitizzate. I risultati mostrano che gli host rock indeformati hanno valori ben concentrati, con ossigeno negativo (da -6 a -4‰ V-PDB) e carbonio positivo (da +1 a +3.5‰ V-PDB). Le vene di dolomite hanno ossigeno leggermente meno negativo dell'host (~ -3.5‰ V-PDB) e carbonio nell'intervallo dell'host (da +0.5 a +3.5‰ V-PDB). Le brecce dolomitizzate hanno l'ossigeno variabile in un intervallo di valori relativamente ampio (da -7.5 a -1.5‰ V-PDB) e il carbonio più costante, al limite superiore dei valori dell'host. Infine, le vene di calcite mostrano la variabilità maggiore, grossomodo organizzata in due andamenti: uno con ossigeno fortemente variabile (da -15 a -4.5‰ V-PDB) e carbonio comparabile all'host, l'altro con ossigeno meno variabile (da -7 a -3‰ V-PDB) e carbonio impoverito (da -7 a +1‰ V-PDB); la variabilità isotopica delle vene di calcite non è associata direttamente alle diverse generazioni (gruppo 1 o 2), nè all'orientazione delle vene, nè alla formazione stratigrafica in cui sono state campionate.

La microtermometria delle inclusioni fluide è stata eseguita su vene sintettoniche di dolomite e calcite. Le vene di dolomite sono caratterizzate da inclusioni bifase a temperatura ambiente che hanno mostrato temperature di omogeneizzazione molto simili tra loro, in un intervallo da 150 a 160°C e temperature di fusione del ghiaccio basse in un intervallo variabile da -18.1 a -12.7°C. Le vene di calcite del "gruppo 1" sono caratterizzate da inclusioni fluidi bifase a temperatura ambiente, ma sono poco adatte all'analisi microtermometrica per via dell'elevata deformazione interna dei cristalli. Ciononostante, sono stati ottenuti risultati attendibili su 4 campioni, che mostrano temperature di omogeneizzazione e fusione del ghiaccio variabili. Le temperature di omogeneizzazione, in particolare, hanno una grande variabilità da un campione all'altro, ma sono relativamente ben concentrate all'interno della singola vena (con picchi localizzati a temperature variabili dagli 80 ai 140°C). Le temperature di fusione del ghiaccio, invece, mostrano una notevole variabilità anche all'interno della singola vena, e nel complesso variano in un intervallo molto ampio (da -20 a -0.5°C). Le vene di calcite del gruppo 2 sono caratterizzate da inclusioni fluide monofase a temperatura ambiente e temperature di fusione del ghiaccio elevate, variabili da -1.3 a -0.3°C.

1.5.4 *Discussione*

Il massimo seppellimento tettonico subito dalla successione Toscana nella zona di interesse è stimato a circa 7 km di profondità e 250°C, corrispondenti ad un gradiente geotermico di 30°C/km (Molli et al., 2011, e riferimenti inclusi). Considerando tale gradiente geotermico, si possono interpretare i dati isotopici e microtermometrici presentati e stimare che le vene di dolomite si siano formate a ~ 175°C e 4.7 km di profondità, per precipitazione da un fluido ad alta salinità (19% NaCl peso equivalente) e composizione isotopica arricchita (~ +13‰ SMOW). Tali caratteristiche sono compatibili con la precipitazione da un fluido di origine locale.

La variabilità di temperatura di omogeneizzazione delle vene di calcite del gruppo 1 è stata interpretata come indicativa della formazione di ogni vena ad una temperatura differente. Considerando nuovamente il gradiente geotermico di 30°C/km, le vene dei campioni analizzati si sarebbero formate a temperature variabili tra i 90 e 160°C, corrispondenti a profondità dai 2 ai 4.5 km. Le temperature di fusione del ghiaccio, molto variabili anche all'interno di ogni singola vena, sono indicative di mixing tra fluidi a bassa ed alta salinità, che possono indicare rispettivamente il coinvolgimento di fluidi di origine superficiale e locale. La firma isotopica del parent fluid, nei quattro campioni vincolati, è arricchita (valori variabili da +6 a +12‰ SMOW) e quindi dominata dal fluido locale. Diverse osservazioni indicano episodi di sovrappressione idraulica durante la formazione delle vene di calcite del gruppo 1, tra cui: i) le relazioni di intersezione non sistematiche tra vene differentemente orientate; ii) episodi di idrofratturazione al footwall dei segmenti di faglia a basso angolo principali, evidenziati dalla presenza di vene corte e dalla forma irregolare; iii) l'assenza di orientazioni preferenziali associate alle successive generazioni di calcite.

Le caratteristiche petrografiche e i dati microtermometrici delle vene di calcite del gruppo 2 sono compatibili con precipitazione a bassa temperatura (< 70°C) da un fluido meteorico a bassa salinità (da 0.5 a 2% NaCl equivalente). La composizione isotopica del parent fluid, pur non essendo vincolata con precisione, è più arricchita delle acque meteoriche e può essere indicativa di interazioni fluido-roccia prima della formazione delle vene.

L'interpretazione dei risultati esposti suggerisce che il Detachment di Tellaro è stato attivo per lungo tempo durante la progressiva esumazione della successione Toscana e la permeabilità associata alla zona di faglia ha permesso abbondante circolazione dei fluidi e precipitazione di dolomite e calcite a temperature e profondità decrescenti. Il sistema di circolazione profondo, durante le fasi incipienti di attività della faglia, sarebbe stato idraulicamente chiuso e dominato da fluidi di origine locale. In queste condizioni si sarebbero formati i corpi dolomitizzati lungo i principali segmenti di faglia a basso an-

golo, che avrebbero costituito una barriera impermeabile e favorito la sovrappressione dei fluidi nella zona di danneggiamento di footwall. Successivamente, durante la deformazione progressiva, i corpi dolo-mitizzati sarebbero stati brecciati e fagliati, mentre il sistema di faglie si sarebbe ingrandito fino a raggiungere la connessione con la superficie. La permeabilità associata al sistema di faglie avrebbe permesso l'infiltrazione di fluidi di origine superficiale nella crosta superiore e il mixing con fluidi più profondi, in un sistema idraulico aperto. Alla fine, quando il sistema di faglie era ormai quasi completamente esumato e ormai inattivo, il sistema di circolazione superficiale sarebbe stato dominato da fluidi di origine superficiale, a bassa salinità.

1.5.5 Conclusioni

Il sistema di faglia del Detachment di Tellaro costituisce un esempio di faglia estensionale attiva durante la progressiva esumazione della porzione interna del prisma di accrezione Appenninico, ed è caratterizzato da: i) contemporanea attività di segmenti di faglia principali a basso angolo e faglie subsidiarie ad alto angolo; ii) elevata pressione dei fluidi e basso stress differenziale. L'ultima condizione, in particolare, avrebbe favorito lo scorrimento lungo i segmenti di faglia a basso angolo, la cui orientazione è meccanicamente sfavorevole.

Il caso di studio del Detachment di Tellaro, quindi, rappresenta un valido contributo ad una tematica tutt'oggi discussa nella comunità scientifica, riguardo la possibile esistenza di faglie estensionali attive ad angoli di immersioni molto bassi, o sulla loro formazione solo in seguito a rotazione passiva di precedenti faglie "Andersoniane".

1.6 LA FOSSA TETTONICA DELLA LUNIGIANA

Questa sezione riassume il capitolo 7. La fossa tettonica della Lunigiana è stata studiata a grande scala mediante l'interpretazione di profili sismici commerciali. L'interpretazione delle principali strutture visibili sui profili sismici e l'integrazione con la geologia di superficie e i dati disponibili in letteratura ha permesso di ricostruire un modello strutturale tridimensionale dell'Appennino Settentrionale, che ha consentito di caratterizzare le strutture a grande scala della catena, fornendo una utile chiave di lettura per interpretare i casi di studio del sistema di faglie della Val di Lima e del Detachment di Tellaro.

1.6.1 Strutture di sottosuolo

I profili sismici analizzati garantiscono una buona copertura del settore Lunigianese e Tosco-Emiliano dell'Appennino Settentrionale e presentano alcuni tratti salienti con caratteristiche analoghe in tutta l'area di studio. In particolare, il dato sismico ha evidenziato la pre-

senza di strutture diverse nei settori interno e centrale della catena, che sono quindi descritti individualmente.

Il settore interno della catena è caratterizzato da un riflettore ben marcato (r), suborizzontale, che si trova generalmente a modesta profondità (3 - 5 km). Tale riflettore è stato perforato, in corrispondenza di un alto strutturale, dal pozzo Pontremoli 1 rilevando la presenza di un basamento cristallino metamorfico. La porzione superiore del prisma, al di sopra del riflettore (r), è caratterizzata da sistemi di faglie estensionali, associati alla morfologia attuale della Val di Vara e dell'alta Lunigiana, a cui sono associate faglie ad alto angolo immergenti verso SW e NE. La porzione inferiore del prisma, al di sotto del riflettore (r), è caratterizzata da strutture a losanga che possono essere interpretate come strutture compressive; tali strutture determinano il sovrascorrimento di scaglie tettoniche chilometriche, a vergenza nordorientale, talvolta caratterizzate da terminazioni in rampa laterale verso NW. Al limite più interno della catena, in prossimità del margine Tirrenico, sono presenti alcuni backthrust SW-vergenti (b₁ - b₃) che tagliano il riflettore (r), determinando complessivamente una separazione di ~ 4 km tra hangingwall e footwall.

Nel settore centrale della catena si osserva l'ulteriore prosecuzione del riflettore (r), qui caratterizzato da un andamento differente: mostra infatti un graduale approfondimento verso NE, fino a raggiungere il limite inferiore del dato sismico. Al di sopra del riflettore (r), la catena è caratterizzata dalla presenza di alcune faglie cieche (t₁ - t₅), che tagliano ad alto angolo i contatti tettonici tra le varie unità tettoniche principali, e determinano il sovrascorrimento di diverse scaglie tettoniche NE vergenti. Le faglie hanno pendenze medie (tra 25 e 40°), le loro terminazioni superiori sono osservate a bassa profondità (tra i 500 e i 1500 m) e determinano rigetti verticali variabili (da alcune centinaia fino a più di 1500 m). La loro estensione laterale è talvolta difficile da vincolare per via della scarsa qualità del dato sismico disponibile. Il sovrascorrimento più esterno (t₅) sembra essere il più esteso lateralmente, coprendo tutta l'area investigata. Il sovrascorrimento più interno (t₁), associato al fronte esposto della successione Toscana non metamorfica, risulta essere il meno esteso lateralmente; è infatti caratterizzato da una terminazione verso NW con una rampa laterale in corrispondenza della finestra tettonica di Pracchiola. In profondità, i rapporti di intersezione fra le faglie inverse e il riflettore (r) sono variabili. In alcuni casi le faglie tagliano il riflettore regionale, determinandone un rigetto rilevante; in altri casi invece i thrust sembrano raccordarsi al riflettore regionale, che costituirebbe il loro livello di scollamento profondo.

Al limite tra il settore interno e centrale della catena, in prossimità dello spartiacque Appenninico, le faglie estensionali immergenti verso SW che bordano l'alta Lunigiana si raccordano con il sovrascorrimento più interno (t₁) ad una profondità di circa 2 - 3 km e determinano una riattivazione in estensione della sua porzione profonda.

1.6.2 *Discussione*

La struttura regionale della porzione analizzata di Appennino Settentrionale è difficile da interpretare univocamente, per vari motivi: i) il prolungamento in profondità, verso NE, del riflettore regionale è ignoto; ii) i rapporti di intersezione fra il riflettore regionale e i sovrascorrimenti che caratterizzano la porzione centrale della catena sono discutibili; iii) la distribuzione della sismicità attuale non rispecchia le strutture visibili nei profili sismici, e terremoti con meccanismi focali estensionali sono presenti nella porzione interna della catena al di sotto del riflettore regionale. L'ambiguità del punto ii, in particolare, non dipende solo dalla naturale variabilità delle strutture geologiche ma anche dalla qualità del dato sismico di sottosuolo: processing differenti della stessa linea sismica, infatti, possono favorire o confutare le varie interpretazioni. Inoltre, la localizzazione ipocentrale dei terremoti è affetta da una significativa incertezza.

Tralasciando il dettaglio di queste problematiche, è comunque possibile estrapolare un modello geologico di prim'ordine, in cui integrare il contesto geodinamico, il dato di sottosuolo, e la geologia di terreno. Tale modello strutturale costituisce una chiave di lettura che permette l'analisi critica dei casi di studio presentati nelle sezioni 1.4 e 1.5. In particolare si può affermare che i sistemi di faglie estensionali ad alto angolo presenti nella porzione interna della catena tagliano una struttura precedente complessa, che comprende sovrascorrimenti, piegamenti a scala chilometrica, e delaminazioni per mezzo di sistemi di faglie estensionali a basso angolo. Inoltre, si può vincolare l'attività del thrust associato al fronte esposto delle unità Toscane non metamorfiche (t_1) ad una fase tettonica intermedia, successiva all'attività dei sistemi estensionali a basso angolo, ma precedente all'attività dei sistemi estensionali ad alto angolo attuali. Questo modello evolutivo permette di conciliare le geometrie interpretabili nei profili sismici e le osservazioni di terreno sul sistema di faglie della Val di Lima e sul Detachment di Tellaro.

1.6.3 *Conclusioni*

Il settore interno del prisma di accrezione Appenninico è caratterizzato da strutture con geometrie differenti nella parte superficiale ($\sim < 5$ km) e in quella sottostante ($\sim > 5$ km). La parte superficiale, in particolare, è caratterizzata da sistemi di faglie estensionali a basso angolo (più vecchi) e ad alto angolo (più recenti). Prendendo spunto dal caso di studio delle Val di Lima, l'attività delle faglie estensionali a basso angolo può essere associata, con una relazione causa-effetto, alle strutture compressive che caratterizzano la parte profonda. La teoria del "critical taper" sull'evoluzione a lungo termine dei prismi di accrezione e il contesto geodinamico permettono, infatti, di ipotizzare che le strutture compressive profonde avessero portato ad un sollevamento

del settore interno del prisma, e innescato l'attività estensionale a basso angolo per compensare il conseguente disequilibrio gravitazionale. Un modello concettuale di prisma di accrezione caratterizzato da una marcata partizione della deformazione può fornire una chiave di lettura per questa fase tettonica. In quest'ottica, il riflettore regionale che caratterizza il sottosuolo dell'Appennino settentrionale avrebbe costituito il "roof thrust" di una struttura a duplex di tipo "antiformal stack", sviluppato in seguito a processi di accrezione basale attivi in profondità.

Successivamente, la catena sarebbe stata caratterizzata da un'evoluzione differente, legata più direttamente alla migrazione del fronte compressivo verso l'avampaese Adriatico e all'instaurarsi di un regime estensionale di retroarco. In questo contesto si sarebbero sviluppate le faglie estensionali ad alto angolo associate alla morfologia dell'Alta Lunigiana e della Val di Vara. Tali faglie ad alto angolo caratterizzano tutt'oggi il quadro sismotettonico dell'area di studio, e proprio la distribuzione ipocentrale della sismicità attuale indica il loro prolungamento in profondità fino a ~ 10 km. Sebbene la loro porzione profonda non sia chiaramente visibile nei profili sismici, si può quindi concludere che tali faglie tagliano, o riattivano in estensione, le strutture compressive più vecchie presenti al di sotto del riflettore r.

1.7 CONCLUSIONI GENERALI

Strutture deformative estensionali sviluppate in contesto regionale contrazionale sono state investigate tramite l'analisi di differenti casi di studio comprendenti strutture a diverse scale, in due differenti prismi di accrezione. Gli esempi selezionati sono ubicati sia al fronte che nelle porzioni più interne dei prismi, ed hanno quindi permesso di investigare strutture associate agli stadi orogenici embrionali e a quelli più evoluti. L'approccio fortemente multidisciplinare ha permesso di evidenziare: i) i rapporti causa-effetto tra deformazioni contrazionali ed estensionali; ii) la complessità dei pattern deformativi che si possono originare in seguito alla ciclica sovrapposizione di deformazioni contrazionali ed estensionali; iii) la permeabilità associata alle strutture estensionali e gli effetti sulla circolazione dei fluidi nella crosta superiore; iv) le implicazioni sul comportamento meccanico dei prismi di accrezione.

I risultati di questo lavoro di ricerca sono di interesse sia accademico che applicativo, con implicazioni che spaziano dai modelli di esumazione delle rocce di alta pressione nelle catene orogeniche, alla possibilità meccanica di attività di faglie misorientate, all'efficace sfruttamento di giacimenti di idrocarburi tramite predizione più accurata della permeabilità per fratturazione in strutture anticlinali.

Part II

ENGLISH VERSION

GENERAL INTRODUCTION

This Ph.D. Thesis deals with the overprinting of extensional and contractional deformations during the development of orogenic wedges. To better understand such a topic, a short introduction about orogenic wedges is provided. The geodynamic context associated with wedge development is briefly introduced and the main geologic processes influencing wedge evolution are characterized. Within this context, some analogue modeling experiments are presented to better characterize those processes and their mutual interplay. Afterward, the topic of extensional deformations developed in contractional regional contexts is addressed by detailed investigations on different case studies.

Selected examples are located in two different orogenic wedges, the Oman Mountains (Sultanate of Oman) and the Northern Apennines (Italy). Selected case studies include structures at different scales and developed at different structural levels. Moreover, selected examples include both “Andersonian” high-angle and misoriented low-angle extensional fault systems.

Case studies are investigated with a multidisciplinary approach that includes: seismic interpretation; analysis of the large-scale structures and investigation of their significance in the regional context; detailed geological and structural mapping; structural analysis at the meso- and micro-scale; petrographic, geochemical, mineralogical and microthermometric analysis of fault rocks and fault-related veins.

2.1 OROGENIC WEDGES

In convergent margins, the movement of one plate toward the other is balanced by subduction. In this process, one plate slips below the other, sinking into the asthenosphere. The subducting plate (the sinking one) is named lower-plate, while the other is named upper-plate. During subduction, different tectonic, gravitational and surface processes characterize the interactions between the upper- and lower-plate. Active processes can be divided in two groups:

- erosive processes, which cause material transfer from the upper- to the lower-plate;
- accretionary processes, which cause material transfer from the lower- to the upper-plate.

Erosive convergent margins develop when erosive processes are dominant. Conversely, accretionary convergent margins develop when accretionary processes are dominant (e.g., Cloos & Shreve, 1988; Von Huene & Lallemand, 1990; Von Huene & Scholl, 1991; Bourgois et al.,

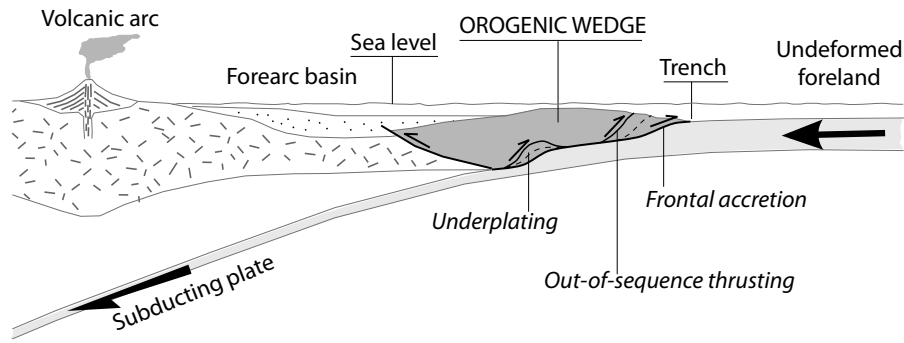


Figure 2.1: Schematic representation of a convergent plate boundary (modified from Dickinson, 1981). Thick continuous lines: active faults; thick dashed lines: inactive faults.

1996; Ranero & von Huene R, 2000; Von Huene, 2003; Clift, 2004; Fisher et al., 2004; Vannucchi et al., 2004; Von Huene et al., 2004; Ranero et al., 2008).

The first-order morphology of a typical accretionary convergent margin includes, from the lower- to the upper-plate: the undeformed foreland, a trench, an orogenic wedge, a forearc sedimentary basin, a volcanic arc (Figure 2.1). The orogenic wedge, in particular, is a wedge-shaped thrust-bounded body, located between the trench and the basement of the upper plate. It is made of materials (portions of oceanic crust and its sedimentary cover) scraped-off from the lower-plate and accreted to the upper-plate (Von Huene & Scholl, 1991). Orogenic wedges are also frequently named “*thrust wedges*”, since they are dominated by thrust faults, or “*accretionary wedges*”, since they result from accretionary processes.

The structure of a typical orogenic wedge is well known to geologists since decades, thanks to the investigation of exposed wedges, subsurface data (mainly provided by industry), and analogue and numerical modeling techniques. The external boundary of the wedge is a frontal thrust which separates the undeformed lower-plate from the deformed units already accreted to the upper-plate. The frontal thrust is characterized by a ramp-and-flat geometry and continues below the wedge with a relatively flat segment. This fault segment, called basal décollement, separates the wedge from the subducting plate and is characterized by intense deformation and important fluid circulation (Moore, 1989).

2.1.1 Main processes

Orogenic wedges are generally characterized by accretionary processes such as *frontal accretion* and *underplating*. Via such processes, materials belonging to the subducting plate are incorporated into the upper-plate in the form of new tectonostratigraphic units. Nevertheless, some processes that dismember the wedge are active as well, and include

basal erosion, surface erosion and tectonic thinning (e.g., Cloos & Shreve, 1988; Von Huene & Scholl, 1991; Lallemand et al., 1994; Dominguez et al., 1998b; Clift, 2004; Vannucchi et al., 2004; Von Huene et al., 2004).

Frontal accretion causes the lengthening of the wedge by accretion of new material at its front. Indeed, with continuous displacement, the deformation front propagates toward the undeformed foreland by nucleation of new thrust faults in more external positions. Each new fault becomes the new temporary frontal thrust of the wedge, while older thrusts may be deactivated or continue their activity as out of sequence (OOS) thrusts (Morley, 1988). It must be noted that frontal accretion does not involve all the material belonging to the lower-plate: a large amount of the subducting materials are not accreted to the wedge, but are dragged below it, in the footwall of the main décollement. This process is named underthrusting.

Underplating, or basal accretion, is active at the base of the wedge, in a more internal position with respect to frontal accretion (e.g., Elliott & Johnson, 1980; Price, 1981; Platt et al., 1985; Hatcher, 1989; Gutscher et al., 1996; Burkhard & Sommaruga, 1998; Mosar, 1999; Kukowski et al., 2002; Avouac, 2003; Malavieille, 2010; Konstantinovskaya & Malavieille, 2011; Long et al., 2011). This process affects some of the underthrust materials and incorporate them at the base of the wedge. Due to underplating, a duplex structure develops at the base of the wedge, and the wedge thickness increases. Wedge thickening is particularly enhanced when the duplex is characterized by a dome-shaped overall geometry; this is the case of antiformal stacking, which causes the uplift of a localized sector of the wedge.

Basal erosion is a tectonic process that allows the transfer of a big amount of material from the upper to the lower plate (e.g., Von Huene & Lallemand, 1990; Von Huene & Scholl, 1991; Bourgois et al., 1996; Von Huene, 2003; Clift, 2004). It takes place at the base of the wedge. In accretionary convergent margins, basal erosion usually affects materials of lower-plate provenance, previously accreted to the wedge; nevertheless, under certain circumstances (e.g., if the wedge is small), also portions of the crystalline basement of the upper plate may be eroded (Von Huene & Lallemand, 1990). As consequence of basal erosion, the whole wedge may undergo important subsidence (e.g., Von Huene & Lallemand, 1990; Clift & MacLeod, 1999; Meschede et al., 1999; Wells, 2003).

Tectonic thinning involves the activity of fault systems that modify the overall shape of orogenic wedges by transferring material from one portion to another (Platt, 1986). A typical example of such processes is represented by low-angle extensional fault systems delaminating the shallow portion of a wedge. Such fault systems can reduce the thickness of a specific portion of the wedge by moving materials to an adjacent portion. Tectonic thinning may not involve material transfer between the upper- and lower-plate, so it does not alter the overall wedge volume.

As a result of accretionary processes, the thickness of the wedge increases and a complex morphology develops, with a large-scale gradient between the wedge and the adjacent trench. Consequently, the superficial portion of the wedge can be affected by surface erosion, that can be active in different forms both in submarine and emerged environments. Then, eroded materials are transported over shorter or longer distances, and deposited in wedge-top basins or in the trench. As a result of the combined action of surface erosion and sedimentation in the trench, materials previously accreted to the wedge are transferred back to the lower plate. Eventually, the same materials may be accreted again to the wedge when the deformation front advances.

2.1.2 *Evolution of orogenic wedges*

Orogenic wedges are usually associated to large-scale convergence and evolve over long time periods (e.g., [Fossen, 2010](#)). All the processes described in the previous section characterize this long-term evolution, but their activity is not constant over time: they mutually influence each other, and they are all affected by the presence of heterogeneities and/or inherited structures in the lower-plate and by changes of the boundary conditions (e.g., the velocity and angle convergent plates, the dip angle of the subducting slab, the climate conditions). As a general rule, all processes show a cyclical and episodic activity characterized by fast- and slow-rate phases.

Consequently, complex deformation structures grow and evolve due to the long-term interactions among the processes that modify wedge dynamics through material transfer ([Elliott, 1976](#); [Chapple, 1978](#); [Storti & McClay, 1995](#); [Mugnier et al., 1997](#); [Konstantinovskaya & Malavieille, 2005](#); [Bonnet et al., 2007](#); [Malavieille, 2010](#); [Perrin et al., 2013](#)). Most of such processes, and their intimate interplay, can be effectively investigated through analogue modeling experiments. In particular, in chapter 3, the evolution of orogenic wedges affected by frontal accretion, basal accretion and surface erosion is investigated. Sintectonic sedimentation and basal erosion have not been taken into account in this experimental program because they had already been exhaustively investigated in previous research (sintectonic sedimentation: e.g., [Cobbold et al., 1989, 1993](#); [Malavieille et al., 1993](#); [Baby et al., 1995](#); [Storti & McClay, 1995](#); [Persson & Sokoutis, 2002](#); [Bigi et al., 2010](#); basal erosion: e.g., [Lallemand et al., 1992, 1994](#); [Malavieille et al., 1992](#); [Mandal et al., 1997](#); [Dominguez et al., 1998a,b, 2000](#); [Nieuwland et al., 2000](#); [Marques & Cobbold, 2002](#); [Hampel et al., 2004](#); [Gutscher et al., 2009](#)). Tectonic thinning has not been investigated because of technical limitations due to the experimental methodology.

From a whole-wedge perspective, the interactions between accretionary and erosive processes is constrained by the mechanical framework of the critical taper theory ([Davis et al., 1983](#); [Dahlen et al.,](#)

1984; Dahlen, 1984, 1990). By the assumption that materials of the wedge are everywhere at yield, such theory relates geometrical and mechanical characteristics of orogenic wedges. In particular, the critical taper theory predicts that wedges tend to attain a mechanically stable surface slope which is related to the strength of the materials constituting the wedge, the dip angle of the basal detachment, and the friction along it. The bigger limitation of this model is that it only gives information about the geometry of the wedge, but it does not give any information about how this geometry is achieved: it does not take into account which tectonic processes occur inside the wedge to attain the stable taper. Nevertheless, the application of critical wedge solutions to fold and thrust belts and orogenic wedges have been among the most significant recent advantages in understanding their large-scale mechanics (e.g., Davis et al., 1983; Dahlen et al., 1984; Dahlen, 1984, 1990; Lallemant et al., 1994). In particular, the application of the critical taper theory to the long-term evolution of orogenic wedges provides a coherent mechanical framework to interpret the interactions between accretionary/sedimentary and extensional/erosive processes. Indeed, accretionary and sedimentary processes increase wedge thickness and re-establish a critical taper when/where the wedge, or a single wedge sector, is in subcritical conditions. Conversely, erosional and extensional processes reduce wedge thickness and re-establish a critical taper when/where the wedge, or a single wedge sector, is in supercritical conditions. Consequently, the long-term evolution of thrust wedges is characterized by the coexistence of contractional and extensional deformations (e.g., Platt, 1986; Burchfiel et al., 1992; Platt, 1993; Wallis et al., 1993; Crespi et al., 1996; Jones & Strachan, 2000; Andresen et al., 2007).

Extensional deformations have long been recognised to characterize the late-tectonic stages of thrust wedge evolution, when compressional deformations attenuate or migrate toward more external sectors (Lister et al., 1984; Malinverno & Ryan, 1986; Dewey et al., 1989; Carmignani & Kligfield, 1990; Patacca et al., 1990; Doglioni, 1991; Burchfiel et al., 1992; Molnar et al., 1993; Carmignani et al., 1994; Crespi et al., 1996; Jolivet et al., 1996, 1998; Reddy et al., 1999; Brogi, 2004, 2008), being triggered by slab rollback and back-arc basin opening (Jolivet et al., 1991, 1994, 1998; Faccenna et al., 1997; Gueguen et al., 1998; Faccenna et al., 2001; Brun & Faccenna, 2008). However, the coexistence of compressional and extensional deformations can also characterize earlier tectonic stages of wedge evolution. Extensional deformations can develop in the outer regions of thrust wedges, associated with: i) the growth of a single fold, being induced by its kinematic evolution; ii) the orogenic history of the whole wedge, being activated to re-establish a critical taper when the wedge is in overcritical conditions (Davis et al., 1983; Platt, 1986), as predicted by the critical taper theory.

2.2 SIN-CONTRACTIONAL EXTENSIONAL DEFORMATIONS

The occurrence of extensional events is common within the evolution of orogenic wedges, as explained in the previous section. The resulting sin-contractional extensional structures can be associated with several processes of scientific and economic significance. Thus, such structures deserve to be investigated in detail.

At a local scale, differently-sized compressional and extensional fault zones can interact during the development of fault-related folds, such as the Jabal Qusaybah anticline described in chapter 4. Indeed, in Jabal Qusaybah, a large-scale anticline is affected by small-scale extensional faults that i) accommodate the fold-related outer-arc longitudinal and transversal extension (i.e. parallel and perpendicular to the fold axis, respectively), and ii) develop in transtensional horse-tail arrays at strike-slip fault tips.

At a larger scale, important stages of extension may result from the overcritical taper conditions of a specific sector of an orogenic wedge, while adjacent sectors might be still characterized by subcritical conditions and compressional deformations. This is the case, for example, of the Val di Lima fault system (chapter 5). In this case study, extensional deformations affect the shallower portion of the Northern Apennines orogenic wedge and are triggered by localized overcritical taper conditions. Additionally, the Val di Lima structural architecture highlights the multiple overprinting of contractional and extensional deformations, cyclically affecting the same area and producing a complex deformation pattern. Indeed, a low-angle extensional fault system occurs in the backlimb of a recumbent fold and is, in turn, affected by open folding and late-orogenic high-angle extensional faulting. The case study of the Tellaro Detachment (chapter 6) is located in a more internal sector of the same orogenic wedge. This area is characterized by well developed extensional deformations, including: i) syn-contractional extensional deformations, that caused a marked thinning of the wedge, and contributed to the exhumation of metamorphic units; ii) late- to post-orogenic extensional fault systems, associated to the migration of the deformation front and the opening of the Tyrrhenian basin.

Finally, the Lunigiana fault system case study (chapter 7) investigates the overprinting between extensional and contractional structures at a regional-scale. This case study includes both syn-orogenic and post-orogenic extensional structures, and provides a regional framework in which is possible to include the Val di Lima and Tellaro Detachment fault systems.

As a general note, it should be pointed out that extensional fault zones influence fluid circulation through the upper crust and modify the mechanical properties of orogenic wedges by favoring fluid overpressure at depths and triggering the development of low-friction hydrous minerals (Hubbert & Rubey, 1959; Davis et al., 1983; Dahlen

et al., 1984; Sibson et al., 1988; Dahlen, 1990; Moore & Vrolijk, 1992; Kirschner & Kennedy, 2001; Collettini & Holdsworth, 2004; Fagereng et al., 2010; Lacroix et al., 2013). Indeed, fault zone architecture includes complex networks of hydraulic conduits, barriers, and mixed conduit-barrier systems (Caine et al., 1996; Storti et al., 2003) that can: i) constitute preferential pathways for fluid migration, allowing the deep infiltration of surface derived fluids (Wickham & Taylor, 1987; Fricke et al., 1992; Morrison, 1994; Morrison & Anderson, 1998; Bebout et al., 2001; Verhaert et al., 2004, 2009; Mulch et al., 2004, 2006; Person et al., 2007; Ingebritsen & Manning, 2010; Gébelin et al., 2011; Gottardi et al., 2011); ii) constitute impermeable barriers favoring fluid compartmentalization at depth and triggering overpressure conditions (Sibson, 1981; Parry & Bruhn, 1990; Sibson, 1992; Gudmundsson, 1999; Sibson, 2000; Gudmundsson, 2001; Collettini & Barchi, 2002; Collettini et al., 2006a; Cobbold et al., 2009). In such circumstances, extensional fault zones can favor mineralizations of remarkable economic significance. Moreover, the contemporaneous activity of extensional and compressional deformations, respectively in the shallow and deep portions of thrust wedges, might favor complex geologic processes such as the sin-contractional exhumation of metamorphic units (Lister et al., 1984; Platt, 1993; Crespi et al., 1996; Jolivet et al., 1996, 1998, 2003; Brun & Faccenna, 2008).

This chapter is presented in the form of a paper published in the *Journal of the Geological Society*. Analog models are used to describe the first-order processes characterizing the long-term evolution of accretionary wedges, and comparison with selected case studies are presented. Analogue experiments presented in this paper allow investigating in detail few key processes that influence wedge evolution such as frontal accretion, underplating and surface erosion. Close attention is paid to the long-term evolution of such processes, their episodic/cyclical behaviour, and the mutual influence of accretionary and erosive processes.

Due to the intrinsic limitations of the experimental procedure, the models presented in this chapter do not directly take into account any sin-contractional extensional deformation. Nevertheless, these experiments highlight some key concepts that allow a better comprehension of the case studies presented in the following chapters.

PUBLISHED ARTICLE

Impact of erosion and décollements on large-scale faulting and folding in orogenic wedges: analogue models and case studies.

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Impact of erosion and décollements on large-scale faulting and folding in orogenic wedges: analogue models and case studies

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Abstract: Deformation mechanisms, long-term kinematics and evolution of fold and thrust belts subjected to erosion are studied through 2D analogue experiments involving large convergence. First-order parameters tested include (1) décollements and/or plastic layers interbedded at different locations within analogue materials and (2) synconvergence surface erosion. Weak layers, depending on their location in the model, favour deformation partitioning characterized by the simultaneous development of underplating domains in the inner part of the wedge (basal accretion) and frontal accretion where the wedge grows forward. Interaction between tectonics and surface processes influences this behaviour. Development of antiformal thrust stacks controlled by underplating shows small- and large-scale cyclicity. Thin plastic layers induce folding processes, which are studied at wedge scale. Recumbent and overturned folds, with large inverted limbs, develop in a shear-induced asymmetric deformation regime via progressive unrolling of synclinal hinges. Surface erosion and underplating at depth induce further rotation (passive tilting) and horizontalization of fold limbs. Model results give insights to discuss the mechanisms responsible for the large-scale structures (i.e. antiformal nappe stacks, klippen and kilometre-scale recumbent fold–nappes) encountered in several mountain belts such as the Montagne Noire (French Massif Central), the Galicia Variscan belt (Spain) and the northern Apennines (Italy).

Supplementary material: Raw data of the experiments are available at www.geolsoc.org.uk/SUP18658.

Orogenic wedges are characterized by complex geological structures growing and evolving over long time periods (e.g. Fossen 2010). Their development is mainly controlled by the general mechanics of subduction and by the interactions between tectonics and surface processes that modify wedge dynamics through material transfer (e.g. Malavieille 2010). Because subduction orogens undergo large convergence, the long-term deformation is intense and generates specific structures, the mechanisms of genesis of which are still not completely understood. Among them, the way in which large-scale recumbent fold–nappes observed in several mountain belt forelands grow and evolve remains enigmatic (Fig. 1). The most beautiful examples in Europe are situated in the Montagne Noire (southern French Massif Central), the Galicia Mountain Belt (Spain) and the northern Apennines (Italy). For example, during the Variscan orogeny, large-scale fold–nappes with huge inverted limbs (sometimes of more than 10 km amplitude) have been created without (or with very low) metamorphism (e.g. Matte 1968; Arthaud 1970). Other large-scale structures are common in mountain belts, such as antiformal stacks of thrust units and subsequent frontal klippen, which can sometimes be related to deformation partitioning and subsequent basal accretion of duplex structures (e.g. Elliott & Johnson 1980; Price 1981; Platt *et al.* 1985; Hatcher 1989; Gutscher *et al.* 1996; Burkhard & Sommaruga 1998; Mosar 1999; Kukowski *et al.* 2002; Avouac 2003; Malavieille 2010; Konstantinovskaia & Malavieille 2011; Long *et al.* 2011; Webb *et al.* 2011). Basal accretion activity is generally not constant during the long-term convergent history of orogenic wedges; underthrusting of

new tectonic slices alternates with the internal deformation of the already accreted ones, or even with the migration of the underplating locus to a new place. Such a cyclical underplating behaviour could promote, at a whole-wedge scale, an alternation from supercritical to subcritical taper condition, which in turn favours the alternations between horizontal and vertical shortening observed in many orogens (e.g. Bell & Johnson, 1989; Bell & Sapkota 2012; Aerden *et al.* 2013). The mechanisms that control large-scale folding, basal accretion and its cyclicity remain an open question.

Analogue modelling is an efficient tool to unravel the main mechanisms controlling the dynamics of orogenic wedges. Various experimental studies have investigated the influence of geometric, kinematic and rheological parameters on the evolution of thrust wedges (see the review by Graveleau *et al.* 2012). One of the most important parameters highlighted is the layering of the accreting crustal materials, which induces mechanical heterogeneity and deformation partitioning. Such a heterogeneous rheology of accreted rock sequences is the result of various factors: stratigraphy of the incoming crustal layer (e.g. Davis & Engelder 1985; Mulugeta 1988; Liu & Dixon 1990; Liu *et al.* 1992; Baby *et al.* 1995; Mandal *et al.* 1997; Nieuwland *et al.* 2000; Costa & Vendeville 2002; Koyi & Vendeville 2003; Konstantinovskaia & Malavieille 2005; Stockmal *et al.* 2007; Malavieille 2010; Smit *et al.* 2010), décollements in a sedimentary sequence or basement–cover interface (e.g. Konstantinovskaia & Malavieille 2011), rheological evolution of the crust owing to *P–T* changes through time (Carry *et al.* 2009; Gueydan *et al.* 2009), or structures and fabrics

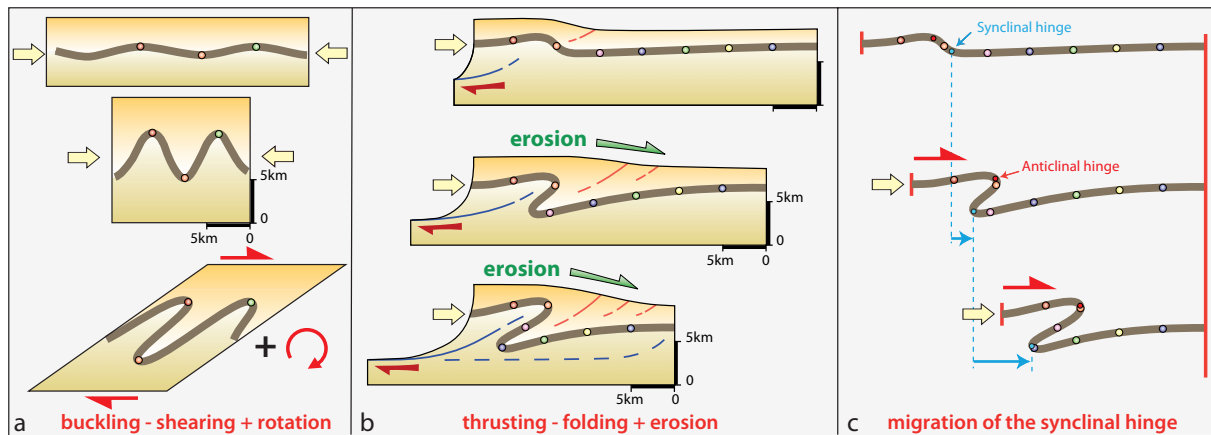


Fig. 1. Schematic illustration showing two folding mechanisms to generate large-scale overturned folds in fold and thrust belts. The first mechanism involves significant burial whereas the second allows the development of large folds with slight burial. (a) Buckling and folding during compressional shortening followed by shearing and tilting of fold limbs. (b) Fold amplification by thrusting and subsequent shearing, unrolling and migration of the synclinal hinge in the frame of a fold and thrust wedge. (c) Simplified kinematic sketch of the same process. Circles on curved line are passive markers regularly spaced in the folded layer.

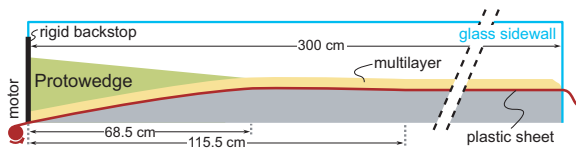


Fig. 2. Experimental set-up. The width of the device is 10 cm.

inherited from an earlier tectonic history (Sutton & Watson 1986; Holdsworth *et al.* 1997; Butler *et al.* 2006, 2008; Bonnet *et al.* 2007, 2008). Among the large number of previous experimental studies of thrust wedges involving multiple décollements, only a few have suggested that fault development and evolution of structures could be cyclical under specific deformation conditions (e.g. Mulugeta & Koyi 1992; Gutscher *et al.* 1998a; Malavieille 2010).

Analogue models allow us to investigate the importance of surface processes (i.e. erosion and sedimentation) and their influence on the dynamics of accretionary wedges (e.g. Baby *et al.* 1995; Larroque *et al.* 1995; Storti & McClay 1995; Mugnier *et al.* 1997; McClay *et al.* 1999; Persson & Sokoutis 2002; McClay *et al.* 2004; McClay & Whitehouse 2004; Konstantinovskaia & Malavieille 2005, 2011; Graveleau & Dominguez 2008; Cruz *et al.* 2008, 2010; Malavieille 2010; Smit *et al.* 2010). Folding represents another aspect of deformation processes investigated using analogue experiments but, although studied at different scales through different experimental setups (e.g. Abbassi & Mancktelow 1990; Grujic & Mancktelow 1995; Tikof & Peterson 1998; Bazalgette & Petit 2007; Noble & Dixon 2011), fold development has rarely been investigated at the scale of a whole accretionary wedge.

This study addresses five major questions, as follows. (1) How do décollements influence deformation partitioning within thrust wedges, and their long-term evolution? (2) Which mechanisms govern the development of large-scale overturned folds commonly preserved in mountain belt forelands as synformal fold-nappe klippen? (3) What controls the growth of large-scale antiformal structures that develop concomitantly in the hinterland? (4) What are the interactions between relatively strong plastic layers and weak décollements? (5) What is the impact of surface processes on

all these mechanisms? We present results of an analogue modelling approach that takes into account large amounts of shortening of mechanically heterogeneous, multilayered materials and simultaneous surface erosion. Our first goal is to analyse the role of strain partitioning in relation to material transfer by erosion. Then, we study the impact of thin plastic layers interlayered with the incoming material on folding mechanisms and on the evolution of folding at the scale of a fold and thrust wedge. The main experimental results are discussed and compared with large-scale tectonic structures from several mountain belts to better interpret their geometry and kinematic evolution.

Experimental set-up and procedure

The experimental set-up simulates the basic geometry and the main mechanisms of a subduction zone where lower plate crustal materials sink beneath an upper plate. This domain of the upper plate located above the subduction interface corresponds in the experiments to a deformable proto-forewedge equivalent to the units of the orogen already accreted, deformed and structured following subduction. All experiments are performed under normal gravity field in a classical sandbox (see Malavieille 1984; Konstantinovskaia & Malavieille 2005, 2011), adapted to allow large shortening (over 200 cm) and showing a flexure of the basal plate that takes into account the curvature of a subducting plate.

The sandbox (Fig. 2) is 10 cm wide and 300 cm long, with a vertical rigid buttress. At the base a thin plastic strip (dacron cloth) exits from the device through a thin slot located at the base of the buttress. It is pulled by a computer-controlled step-by-step electric motor. Analogue materials representing the upper crust rocks of the lower plate are deposited onto the plastic strip and are dragged toward the buttress. As they cannot exit from the device, they are accreted against the upper-plate backstop. A thin layer of sand is glued on the upper surface of the plastic strip to form a very rough surface. This creates a high basal friction ($\mu_b \approx 0.5$) between the basal strip and the analogue material of the models. According to the critical wedge theory (Davis *et al.* 1983; Dahlen 1984; Dahlen *et al.* 1984), the strength of the basal décollement influences the dip of the main thrusts and backthrusts and the surface slope angle of a wedge that satisfies the yield conditions.

Three materials are used in the models. (1) Aeolian sand, with a density of 1690 kg m^{-3} , well-rounded grains less than $300 \mu\text{m}$ in size, a coefficient of internal friction (μ_0) of 0.57 and cohesion (C) of 100–150 Pa, forms the upper plate protowedge and a large part of the lower plate layers. (2) Glass microbeads poured into the sandcake are used to model weak layers (décollements). The grain diameter is 100–200 μm , and the perfect roundness of the grains leads to a smaller coefficient of internal friction than aeolian sand ($\mu_0 = 0.44$) and a negligible cohesion. (3) Plasticine is used to simulate folding because it shows plastic behaviour. It is composed of mineral oils, waxes and a solid filler made of fine powder (15 μm). It does not contain water, does not dry, and can be reused. The plasticine is melted in an oven (softening point between 39 and 42°C) and poured into a $200 \text{ cm} \times 9.8 \text{ cm}$ mould. After cooling, 1 mm thin layers are sliced and then included in the multilayered model. To observe the final deformation of the plastic layer at the end of experiment, we carefully cleaned the sand around the fold limbs. Practically, in the experimental procedure, when a plasticine layer is emplaced in the sand cake, the width of the thin plasticine sheet is slightly less than that between the two glass sidewalls to avoid parasitic effects of lateral friction along the glass. Therefore, less than 1 mm of sand separates the plastic layer from the sidewall. As a consequence, what we can observe directly through the glass sidewall is not the plasticine layer itself, but the thin coloured sand marker that outlines its deformation.

Aeolian sand and glass microbeads are commonly used in physical modelling studies as analogue of upper crustal rocks with a brittle behaviour. The scaling factor between their mechanical properties and those of the natural prototype is 10^5 (Krantz 1991; Schellart 2000; Lohrman *et al.* 2003). The same 10^5 scaling factor is therefore used for the model dimensions (1 cm = 1 km), to satisfy the fundamental scaling theory for analogue modelling (Hubbert 1937, 1951; Horsfield 1977; Ramberg 1981; Davy & Cobbold 1991; Graveleau *et al.* 2011). Plasticine is a non-Newtonian fluid characterized by strain rate-dependent plastic yielding and strain hardening. At constant temperature, the constitutive flow law for plasticine is given by $\dot{\epsilon} = C\sigma^n$, where $\dot{\epsilon}$ is the strain rate, C is a material constant, n is a stress exponent, and σ is the differential stress (McClay 1976; Ranalli 1995). The apparent dynamic viscosity (η) of plasticine is given by one-half the ratio between the differential stress and the strain rate: $\eta = \sigma / (2\dot{\epsilon})$.

Considering that the experiments were carried at constant room temperature ($T \approx 22\text{--}25^\circ\text{C}$) and that the strain rate of the plasticine layer during deformation is constant and very low ($\dot{\epsilon} \approx 1 \times 10^{-3} \text{ s}^{-1}$) then $\eta \approx 4 \times 10^7 \text{ Pas}$. This value of viscosity was determined by mechanical experiments on plasticines (Schöpfer & Zulauf 2002; Zulauf & Zulauf 2004) whose composition is similar to that of the plasticine used in our analogue models.

The yield strength of a plasticine layer with millimetre thickness is roughly equivalent to the compressional strength of a sand layer with centimetre thickness. Thus, the strength contrast between plasticine layers and granular layers (sand or glass microbeads) in our models is roughly equivalent to the strength contrast between ductile yet strong rock layers (such as limestone) and weak rock layers (such as siltstones or shale). This strength contrast is typical of sedimentary sequences observed in foreland and intramontane basins in orogenic wedges.

Thirteen experiments have been run (Table 1). Of these, seven have been chosen as representative to describe the main results of our study. Erosion has been applied to most of them, using the following procedure. First, an initial shortening without erosion is applied to the models, allowing the development of a wedge-shaped topographic relief. This first step of wedge growth could be considered as the analogue of wedge development in a slightly erosional

submarine setting. Then, erosion is applied step by step, each 2 cm of convergence (see digital screen in experiment figures for shortening values) simulating climate-dependent erosion in a subaerial wedge setting and maintaining a constant evolution of the wedge topography. This is performed by scraping off all the material rising above an imposed erosion surface and then removing it with a vacuum cleaner. The slope of this erosion surface can be variable or fixed and predetermined. In the former case the slope of the erosion surface is adjusted step by step, to follow the ‘instantaneous’ average slope of the wedge, smoothing in this way only the small-scale irregularities of the topographic profile without altering its average slope. In the latter case an average tilted erosion profile dipping from 3 to 10° toward the foreland is maintained (Konstantinovskaia & Malavieille 2005, 2011; Bonnet *et al.* 2007, 2008). This slope corresponds to the critical taper slope of a dry sand wedge (Davis *et al.* 1983). In this case, local erosion rates are directly controlled by the activity of thrusts. Although the role of sedimentation has not specifically been taken into account in this study, small piggyback basins that develop during wedge growth have been filled while performing erosion to avoid unrealistic foreland topographies.

Experimental results

Our new set of experiments complements previous modelling studies on orogenic processes carried out at the Geosciences Laboratory in Montpellier, which outline the impact of coupling between surface and tectonic processes and the important role of décollement levels during deformation (see Malavieille *et al.* 1993; Larroque *et al.* 1995; Konstantinovskaia & Malavieille 2005, 2011; Bonnet *et al.* 2007, 2008; Malavieille 2010). Décollement layers favour the mechanical decoupling of stratified material of the subducting plate and consequently induce deformation partitioning. Whereas upper units are accreted at the toe of the wedge during propagation of the deformation front (frontal accretion), lower units are underthrust below the main décollement fault and accreted at the base of the wedge by duplexing and underplating (i.e. basal accretion). This partition between vertical and horizontal accretion has a major impact on the organization of tectonic structures, deformation and exhumation of deep units (e.g. Gutscher *et al.* 1998a,b; Bonini 2001, 2003; Adam *et al.* 2002; Kukowski *et al.* 2002; Konstantinovskaia & Malavieille 2005, 2011; Hoth *et al.* 2006, 2007, 2008; Bonnet *et al.* 2007, 2008; Malavieille 2010).

Influence of weak layers

A microbeads layer deposited on the top surface of the lower plate (experiment 1, Fig. 3a) favours underthrusting of the tectonic units below the protowedge in the first steps of experiments. During shortening, the wedge grows mostly in sequence by frontal accretion and shows the typical structure of a simple sand wedge with no décollement in the incoming sequence. We recognize typical structures of high basal friction wedges built up by underthrusting of long tectonic units (e.g. Malavieille *et al.* 1992; Lallemand *et al.* 1994; Gutscher *et al.* 1998a,b; Nieuwland *et al.* 2000; Agarwal & Agrawal 2002; Kukowski *et al.* 2002; Konstantinovskaia & Malavieille 2005; Graveleau *et al.* 2012). In some cases, when the deformation front advances toward the foreland through the nucleation of a new thrust, the previous frontal thrust remains active until the end of the experiment. Internal deformation of single forward vergent thrust units is accommodated by small backthrusts and only few large backthrusts propagate through the whole wedge.

A microbeads layer positioned at about one-third of the total lower plate thickness favours remarkable strain partitioning (experiment 4, Fig. 3b). Materials of the upper portion of the subducting

Table 1. *Parameters used for the 13 experiments*

Experiment number	Total shortening (cm)	Protowedge		Lower plate		D��collement position* (cm)	Plasticine			Erosion		
		Length (cm)	Slope (deg.)	Total thickness (cm)	Termination		Present	Thickness (cm)	Position* (cm)	Present	Start† (cm)	Slope
1	150	61	7	3.1	Straight	Top lower plate	No			No		
2	150	66	4	3.1	Straight	Top lower plate	No			No		
3	160	80	2	3.1	Gradual	Top lower plate	No			No		
4	180	74	0	3.1	Straight	1.1	No			No		
5	201	74	0	3.1	Straight	1.1	No			Yes	30	Variable
6	198	66	4	3.1	Straight	2.1	No			Yes	30	Variable
7	220	66	4	3.1	Straight	2.1	No			Yes	20	Fixed (5��)
8	160	80	2	3.3	Gradual	Top lower plate	Yes	0.3	1.7	No		
9	167	90	0	3.3	Gradual (��15 cm)‡	Top lower plate	Yes	0.2	1	No		
10	163	90	0	3.3	Gradual (��15 cm)‡	Top lower plate	Yes	0.1	1	No		
11	193	90	0	3.3	Gradual (��15 cm)‡	Top lower plate	Yes	0.1	1	Yes	25	Fixed (5��)
12	175	80	2	3.3	Gradual	1.3	Yes	0.2	1.5	No		
13	200	90	0	3.5	Gradual (��15 cm)‡	1	Yes	0.1	1.7	Yes	15	Fixed (3��)

Models described in the text are shown in bold type.

*Height from base plate.

†Amount of initial shortening without erosion.

‡Lower plate ends before the backstop.

plate are deformed by frontal accretion, leading to the development of a typical low basal friction thrust wedge. Given the reduced thickness of the materials deformed by frontal accretion and the reduced basal friction, both the wedge taper angle (Fig. 3b) and the spacing between new thrust faults are smaller than in experiment 1. Materials of the lower portion of the subducting plate are deformed by underplating and basal accretion (Fig. 3b). As already described in previous papers (e.g. Mulugeta & Koyi 1992; Gutscher *et al.* 1996, 1998a), underplating is not a steady-state process. More in detail, it can be noted that the evolution of the duplex structure is characterized by the activity of (1) large-offset faults (continuous lines in Fig. 3b and c), allowing the accretion of a new unit at the base of the wedge, and (2) small-offset faults (dotted lines in Fig. 3b and c), allowing the accommodation of the internal deformation of single underplated units (Adam *et al.* 2002; Kukowski *et al.* 2002; Hoth *et al.* 2008). The large shortening of the experiments presented in this paper also allowed a second type of cyclicity to be observed at the whole-wedge scale. When the first antiformal stack of underplated units reaches a critical size, it becomes inactive and the locus of underplating shifts to a more external position (e.g. toward the foreland), where a second antiformal stack starts developing. The first underplating domain is thus passively accreted and becomes part of the wedge upper plate. Such a mechanism of accretion is repeated cyclically during the long-term evolution of a wedge.

Growth of the antiformal stacks leads to some localized uplift of the overlying portion of the wedge as manifested by the shape of the topographic profile, which shows two bumps corresponding to the deep duplexes. In contrast, the topographic profile is regular in wedges lacking strain partitioning (Fig. 3a and b).

A microbeads layer located at shallow depth in the sandcake has a slight influence on active deformation, without leading to efficient strain partitioning (experiment 6, Fig. 3d). Only second-order small thrust units develop along the weak layer, involving the upper portion of the subducting plate. Such small thrusts alternate regularly with major thrusts involving the whole subducting plate, and can therefore be considered local splays of the main thrusts. The overall architecture of the wedge is that of a classical high basal friction wedge, similar to the one obtained without a microbeads layer (Fig. 3d).

Influence of a plastic layer on folding

In experiment 10 (Fig. 4a) the introduction of a plastic layer in the subducting plate drastically changes the tectonic style and kinematic evolution of the wedge, and leads to folding processes at the scale of a whole accretionary wedge. The main mechanisms of fold development in the experiment with no erosion are described in Figure 5. Development of folds begins by buckling of the plastic layer, which is accommodated by a pop-up structure in the overlying ‘brittle’ sand layer. The wavelength of buckling and the folding mechanism are controlled by the strength contrast between the relatively strong yet ductile plasticine layer and weaker sand layers (see the previous section). Soon afterwards, the folding amplitude increases, and folds become asymmetric and overturned, controlled by progressive shearing deformation induced by the growth of the prism. Shearing is partly responsible for the development of the long inverted limbs, but the plastic layer is never disrupted or stretched enough to be cut, thus allowing unrolling of the synclinal

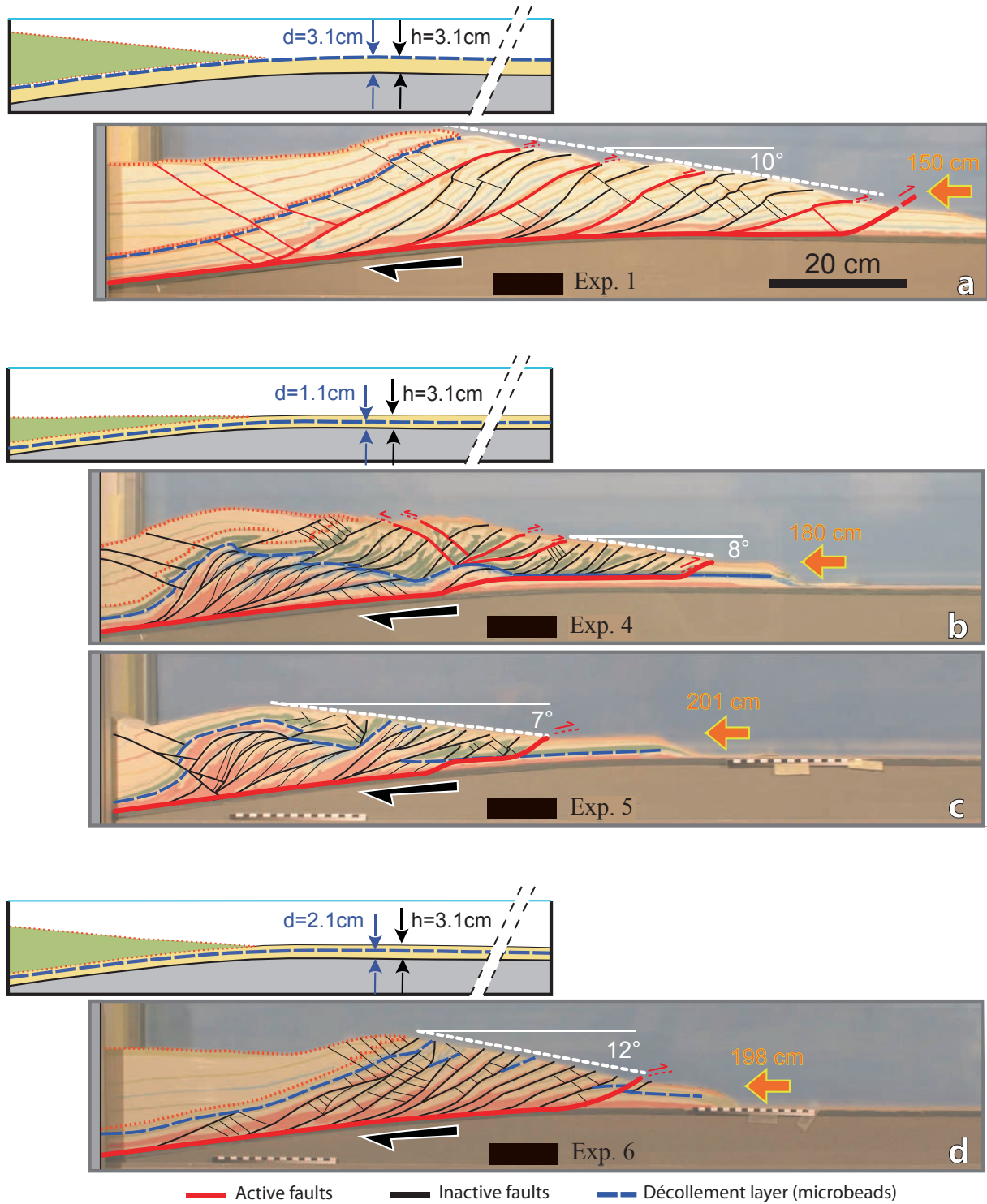


Fig. 3. Initial configuration and final result of models involving a microbeads layer located at different heights d . (a) Classical high-friction thrust wedge. The thrust front propagates in a piggyback style, but some faults remain active after the nucleation of new faults at the front. (b) Strain partitioning: duplexing at the base of the wedge and frontal accretion at the toe. The cyclical behaviour of underplating should be noted. (c) Impact of surface erosion on the wedge dynamics, location of underplating and exhumation of underplated units. (d) Underplating is inhibited when décollement is too shallow. Shortening is indicated above the orange arrows.

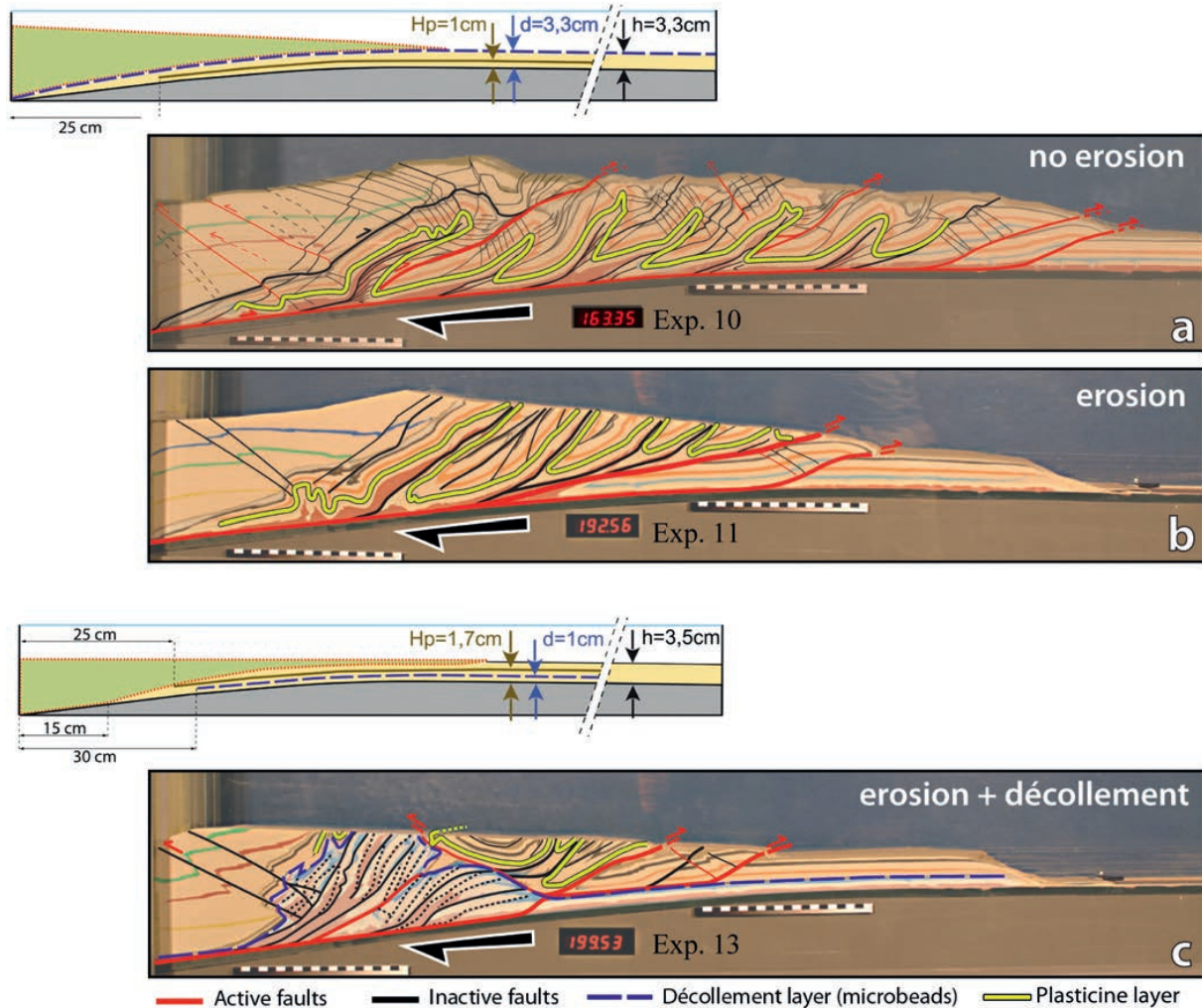


Fig. 4. Initial setting and final stage of model involving a thin layer of plasticine (H_p , height of the plasticine layer in sketch). (a) Case with no erosion. Folds are overturned at an angle between 30 and 50°. (b) Same experiment with erosion. (c) Brittle-plastic model with décollement layer and erosion. Shortening is given in centimetres on digital screen.

hinge. Fold growth proceeds by continuous unrolling of the synclinal hinge, which causes existing nearby thrusts to become inactive while new ones form. Then, the inactive faults are passively transported along the inverted fold limb, although some can be reactivated by out-of-sequence thrusting.

Impact of surface erosion

The effect of surface erosion is tested on wedges characterized by different tectonic styles. In experiment 6, a simple high-friction thrust wedge showing no strain partitioning or folding is subjected to erosion. It is comparable with experiment 1, except that the latter did not include erosion. Figure 3d outlines the similar tectonic styles of both models, and similar taper angle of their pro-wedges.

Experiment 5 (Fig. 3c) shows the effects of erosion on a wedge characterized by strong strain partitioning, to be compared with experiment 4, which has the same initial setup but no erosion. In both models permanent underplating leads to the development of a large antiformal stack formed by basal accretion of duplex

units. It induces the rise of the internal domain and subsequent localized surface uplift. In model 5, however, surface uplift enhances localized surface erosion, which in return favours further uplift and localization of underplating. As shown in Figure 3c the final product of this process is the exhumation of under-plated units in localized areas.

In experiment 11 (Fig. 4b) a model involving a plastic layer is subjected to erosion to investigate the impact of surface processes on the dynamics of folding. Compared with experiment 10, which has the same initial setup, experiment 11 shows several differences in the general tectonic style. Despite erosion of anticlinal fold hinges, shear deformation increases the length of the inverted sequence, which is associated with the unrolling and migration of the synclinal fold hinge. Frontal accretion spreads but the amount of shortening accommodated through each folded tectonic unit is higher. The material removed by erosion delays the growth of the wedge and thus its ability to propagate the deformation forward through new tectonic units. Although the anticline hinge is being removed by erosion, unrolling of the syncline hinge continues, suggesting that the inverted limb is

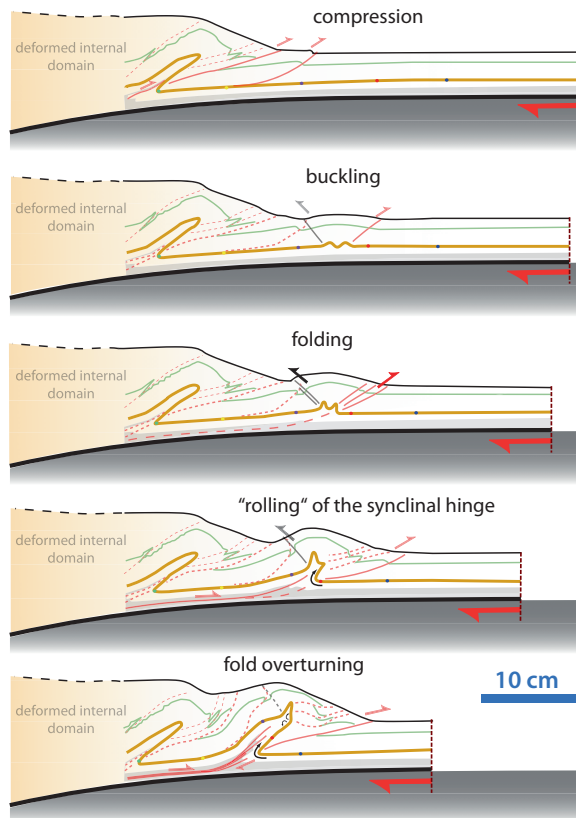


Fig. 5. Kinematic model of folding. The curved line with circles represents the plasticine with displacement indicators.

not subjected to traction. The diffuse shear deformation involved in the core of folds owing to asymmetric shortening could be responsible for the forward migration (relative to the undeformed foreland) of the synclinal hinge.

To summarize the main differences, we note that each unit is more intensely deformed and that the length of fold limbs is greater in the experiment with erosion. In addition, for an equivalent amount of shortening, fewer tectonic units were formed.

Brittle–ductile multilayer and underplating

In experiments 10 and 11, large-scale isoclinal folds developed with an average final overturning of the limbs ranging between 30 and 50°. These modelling results cannot explain what is commonly observed in many natural mountain forelands, where large-scale inverted fold limbs are close to horizontal over kilometres. Thus, important questions remain: what mechanism is responsible for the huge overturning observed? Also, does this mechanism occur during fold development or by rigid rotation owing to late tilting? To answer these questions, we have taken into account the insights from the experiments involving basal accretion. As previously shown, décollement layers play a key role during deformation and interactions with surface processes, which seems to be important also for the development of folding during the growth of fold and thrust belts.

In experiment 13 (Fig. 4c) we tested the impact of heterogeneous layering involving décollements, and brittle and plastic behaviours

(very common in foreland belts). The model combines strain partitioning, folding and surface erosion. A 5 mm thick layer of sand is placed between a 1 mm thick plasticine sheet and the 3 mm thick weak layer of glass microbeads. The complete evolution of the experiment is described in Figure 6. Erosion begins after 15 cm of shortening and the wedge slope is sustained at about 3° during shortening (60% at the end of the experiment). During convergence, fold hinges are rapidly eroded, and active thrusting occurs in the core of folds. Six folded tectonic units are obtained and a large domain of deeply accreted units is exhumed behind the prism. It should be noted that we also observed a cyclical behaviour of underplating as described above.

The structures located above the décollement layer are passively deformed and uplifted owing to basal accretion, tilting the back part of folded tectonic units. These deep accretionary processes are responsible for the important overturning and rotation of the flanks of folded structures. During continuous shortening, the kinematics of deformation reflects the complex interaction between wedge mechanics and erosion. At the final stage most of the folded units from the back part of the wedge have been removed by erosion, and three tectonic domains characterized by specific deformation features are juxtaposed. From the frontal part of the wedge to the back-stop, we have (Fig. 6) a frontal imbricate of thrust and fold sheets, a synformal klippe of folded units previously accreted to the front and progressively deformed, and an antiformal stack of underplated thrust units refolding the upper décollement layer.

Discussion and case studies

Results of this series of experiments give some insights for the interpretation of several debated features of the forewedge domain of mountain belts formed in continental subduction settings. Chosen case studies are discussed in the light of our experimental results. The last model, which contains the main features described in previous sections, is used to illustrate the general mechanisms explaining the relationships between the main tectonic units of natural orogenic wedges (Fig. 7).

Examples from the Variscan Belt

The Variscan orogen developed during the Gondwana–Laurasia collision from Devonian to middle Carboniferous times (e.g. Matte 2007). The Montagne Noire in the southern French Massif Central and the Galicia Mountains in NW Spain represent segments of this orogen characterized by a foreland fold and thrust belt domain associated with a syntectonic foreland basin (e.g. Matte 1968; Arthaud 1970; Pérez-Estàun *et al.* 1991; Simancas *et al.* 2013). Low-grade tectonic units mainly composed of sedimentary rocks of the Palaeozoic cover (schists, limestones and quartzites of Cambrian to Carboniferous age) are intensely folded and juxtaposed with antiformal stacks of Proterozoic to Cambrian metamorphic basement rock units largely exhumed in the hinterland. The Montagne Noire, which forms the southernmost part of the Variscan French Massif Central (Fig. 8), is generally subdivided into three tectonostratigraphic units (e.g. Gèze 1949; Arthaud 1970). (1) A Northern Flank upper-plate unit with a southward tectonic vergence consists of folded and faulted low-grade lower Palaeozoic metasedimentary rocks. (2) An Axial Zone lower-plate unit is formed by an antiformal structure of crystalline rocks (gneiss, migmatite, and micaschist) of Proterozoic to Ordovician age. This metamorphic domain composed of high-grade rocks has been variously interpreted in terms of diapirism (e.g. Gèze 1949; Beaud 1985; Charles *et al.* 2009), contractional tectonics (Arthaud *et al.* 1966; Mattauer *et al.* 1996; Aerden & Malavieille, 1999; Soula

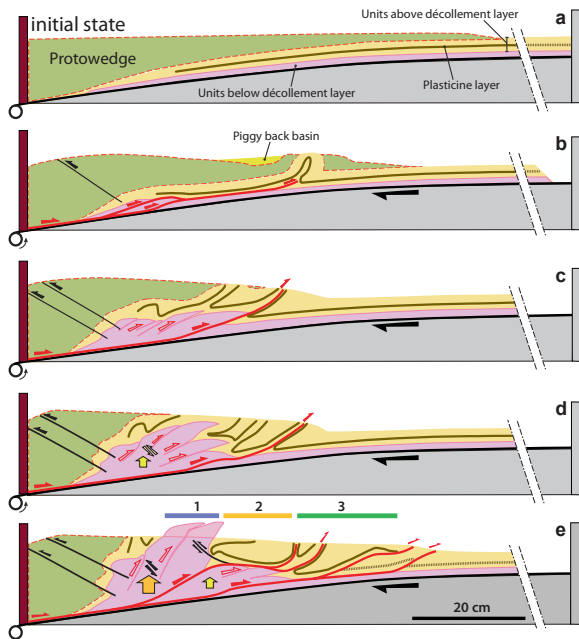


Fig. 6. Evolution of complex model involving brittle-plastic multilayer, erosion and décollement layer. Upper units situated above décollement layer are composed of sand and plasticine; deep units below décollement layer are composed of sand. At the final stage (e), three domains are juxtaposed: (1) exhumed deep rocks; (2) recumbent folds; (3) overturned folds.

et al. 2001; Matte, 2007; Malavieille 2010), emplacement in a crustal-scale strike-slip setting (e.g. Nicolas *et al.* 1977; Franke *et al.* 2011), or as an extensional metamorphic core complex (e.g. Echtler & Malavieille, 1990; Van den Driessche & Brun 1992). In fact, most researchers agree on the geological evidence for a contractional history followed (or assisted) by gravity-induced extensional processes favouring exhumation, detachment formation and diapirism in the evolutionary stages of the orogen. (3) A Southern Flank, well known in the literature for its kilometre-scale recumbent fold-nappes, is composed of very low-grade Palaeozoic sedimentary sequences. The south-verging nappe stack is intimately associated with syntectonic Visean flysch sediments deposited in a foreland basin setting, in a shallow marine environment. The upper-plate nappes are separated from high-grade lower-plate basement units by major fault zones that record a complex pattern of deformation (e.g. Echtler & Malavieille 1990; Aerden & Malavieille 1999). Figure 8 shows an interpreted cross-section of the Montagne Noire (Malavieille 2010).

Similar key structures of the Galicia Mountain Belt are outlined on the cross-section shown in Figure 9, modified from Pérez-Estàun *et al.* (1991). Proterozoic metamorphic units crop out in the internal domain whereas a domain of large-scale recumbent folds made of Cambrian quartzite or limestone characterizes external foreland units. These kilometre-scale folds have horizontal or overtilted limbs (Matte 1968; Pérez-Estàun *et al.* 1991). It should be noted that in this segment of the Variscan belt, a cyclical basal accretion may have occurred at large scale, as two antiformal stack structures formed during wedge growth.

Comparing wedge architecture and analogue models, we can outline geometrical and kinematic similarities. As observed in models

involving décollements and brittle-plastic behaviour, there is a good analogy between the geometric configuration of the folded superficial domain and the underlying deeper structures. Where underplating develops, the folded units located above the décollement layer are strongly tilted by subsequent uplift in the antiformal stack domain. This can be compared with the fold-nappe structures of the Galicia belt or Montagne Noire. In the parts of the orogenic wedge located far from the locus of basal accretion, large-scale folds are simply overturned. During the growth of the Montagne Noire, syntectonic flyschs (Visean) are deposited at the toe of the wedge owing to erosion of developing fold-nappes (Southern Flank). At the same time, basal accretion is active during the growth of the prism involving underthrusting of sliced Proterozoic basement and subsequent uplift at the back of the overturned fold domain. Already deformed Upper Palaeozoic units are overtilted by progressive uplift. The décollement layer allowing strain partitioning between shallow and deep parts of the wedge is located along the main inherited discontinuity, between the crystalline basement and the Palaeozoic cover.

Figure 6 highlights the major effect of erosion on deformation processes. During wedge growth, as a result of combined basal accretion, surface uplift and erosion, continuous folding affects tectonic units of the upper plate, which remain at the same structural level in the upper crust, whereas large domains of deep metamorphic units of the lower plate are exhumed. Thus, the deformation mechanisms highlighted in our study may explain how the large-scale recumbent fold-nappes with inverted limbs of 10 km develop, and why they underwent only slight or no metamorphism.

Northern Apennines

The Apennines are a fold and thrust mountain chain forming the backbone of the Italian peninsula. Figure 10a outlines some aspects of the northernmost portion of this chain (e.g. Molli 2008). The internal zone is characterized by a metamorphic core where two main exhumed tectonostratigraphic units crop out. The lowermost is the Apuane unit, a low-grade metamorphic unit showing greenschist assemblages (deepest estimated burial *c.* 20 km). On the western side of the core, the Apuane unit is overlain by the Massa unit, an HP greenschist-facies metamorphic unit with higher grade *P-T* peak conditions (estimated deepest burial *c.* 25–30 km). The whole metamorphic core is overlain by the Tuscan Nappe, an anchimetamorphic unit with a deepest estimated burial of *c.* 7 km, which in turn is overlain by Subligurian and Ligurian non-metamorphic units and by the Epiligurian basin (e.g. Fellin *et al.* 2007, and references therein). In the central portion, east of the Alpi Apuane, the chain is characterized by a recumbent fold domain (Fig. 10b) where the Tuscan Nappe is folded in a kilometre-scale recumbent structure (the Val di Lima fold), with an outcropping kilometres-long inverted limb. This recumbent fold shows minor structures related to superimposed deformations (Baldacci *et al.* 1992; Fazzuoli *et al.* 1998) with development of the long inverted limb by progressive hinge migration (Botti *et al.* 2010).

The overall geometrical configuration of the analysed segment of the Apennines (Fig. 10) can be interpreted in the light of our models. The combined action of basal accretion and underplating of tectonic units produced the growth of a synmetamorphic antiformal stack (Molli & Vaselli 2006) responsible for strong uplift and exhumation by submarine (in the early stages) and later surface erosion (Molli *et al.* 2002; Fellin *et al.* 2007). Erosion processes were associated with tectonic thinning by normal-slip reactivation of the basal thrust of the unmetamorphic units at the hanging wall and the metamorphic core at the footwall (Carmignani & Kliffeld 1990; Molli *et al.*

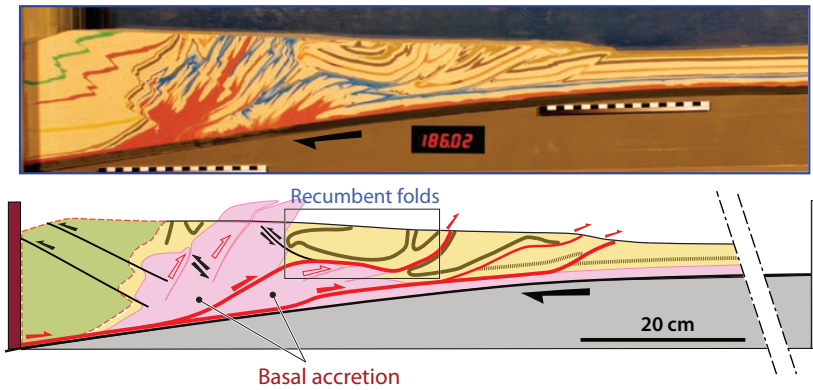


Fig. 7. Relationships between cyclical duplexing and kinematics of folding in a brittle-plastic model with erosion and décollement level.

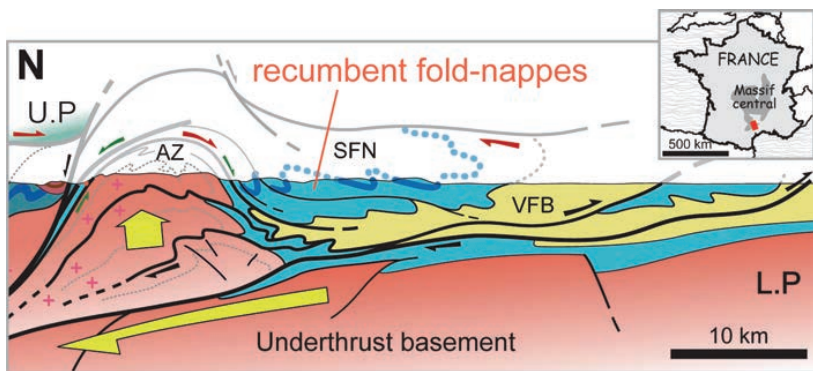


Fig. 8. Cross-section of the Montagne Noire, southern part of French Massif Central (modified from Malavieille 2010). U.P., upper plate; L.P., lower plate; AZ, axial zone; SFN, southern fold-nappes; VFB, Viséan foreland basin.

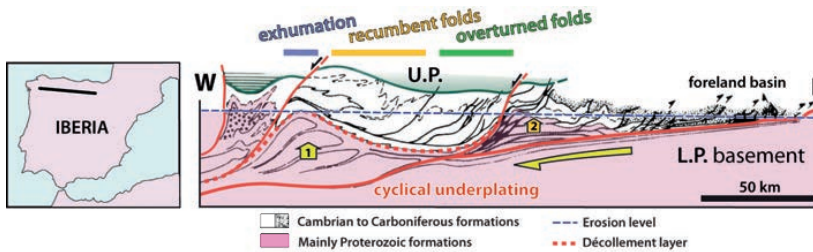


Fig. 9. Cross-section of Variscan Belt in Galicia, Spain (modified from Pérez-Estáun *et al.* 1991). U.P., upper plate; L.P., lower plate. Arrows 1 and 2 are locations of distinct basal accretions.

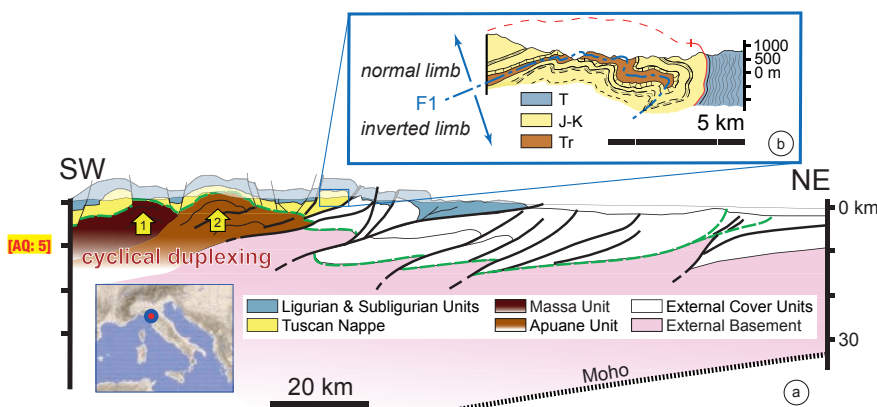


Fig. 10

Fig. 10. (a) Cross-section of the northern Apennines, Italy (modified from Molli 2008). (b) Cross-section of the Val di Lima fold structure (modified from Baldacci *et al.* 1992). T, Tertiary; J-K, Jurassic-Cretaceous; Tr, Triassic. Arrows 1 and 2 are locations of basal accretions.

2002; Fellin *et al.* 2007; Molli 2008). Moreover, the growth of the antiformal stack may have induced progressive tilting of the basal detachment and the formation of recumbent geometry of the folded Tuscan unit in the Lima Valley (Fig. 10a and b).

Indeed, the locus of basal accretion and the folded domain are close enough to presume the influence of the underlying deep structures on the passive rotation of the fold limbs. Model results can give some more hints on the processes that may have influenced the geological evolution of the metamorphic core, suggesting that two successive episodes of underplating could have been responsible for the syncontractional juxtaposition of the Massa unit above the Apuan unit.

Open questions in other mountain belts

Mechanisms responsible for deformation structures and exhumation processes developed in the Himalayas are at present widely discussed through two main types of models: ductile channel flow (e.g. Nelson *et al.* 1996; Beaumont *et al.* 2001; Jamieson *et al.* 2004) and wedge extrusion in a thrust system (e.g. Burchfiel & Royden 1985; England & Molnar 1993; Guillot & Allemand 2002; Webb *et al.* 2007; Kali *et al.* 2010). Our study outlines simple mechanisms that seem to be consistent with the observed large-scale geological structures (antiformal stacks, synformal klippen of fold and thrust units), in agreement with the second orogenic wedge model.

Conclusions

Interaction between climate-controlled surface processes including erosion, sedimentation and deformation processes plays a key role in the structural evolution, kinematics and exhumation of rocks in orogenic wedges. During continental subduction, the role of the rheological layering of the crust can be major, as it determines the partitioning of deformation in a growing orogenic wedge into domains undergoing horizontal and vertical accretion. Partitioning is first controlled by tectonic processes, but material transfer induced by surface processes exerts a strong feedback on wedge dynamics. Insights from analogue models applied to natural cases allow us to emphasize several first-order interaction mechanisms that result from this coupling. Experiments show that strain partitioning is not systematic but depends on the position of weak layers in the layered incoming sequence. They show a cyclical behaviour of basal accretion, leading to episodic underplating of tectonic units, which has a strong impact on the vertical component of displacement of rock material. In turn, this changes surface slopes, favouring erosion in domains of strong surface uplift. In addition, our experiments offer an explanation for the enigmatic domains of non-metamorphic large-scale fold-nappe units observed in the foreland of many orogenic wedges. To a first order, the dynamics of folding involves rolling of a synclinal hinge and develops exclusively overturned fold types. This mechanism was observed, in particular, for multilayered models consisting of both strong yet ductile layers and comparatively weaker granular layers.

The influence of the deep wedge dynamics, such as the growth of basal duplexes, causes further rotation of fold structures, leading to the horizontalization of fold limbs, while erosion processes keep the folded units in a superficial low-grade metamorphic domain. Natural wedges (e.g. Galicia, Montagne Noire and northern Apennines) show close similarities to the experiments described herein, both in terms of architecture and orogenic dynamics. Other orogenic wedges exposing similar structures, such as exhumed antiformal metamorphic domes juxtaposed with domains of largely folded upper-crustal rock sequences, need to be revisited in the light of the general mechanisms here outlined.

Future work should concentrate on multilayered models with different rheological contrasts (e.g. weak ductile layers and strong brittle layers) to determine other possible large-scale folding mechanisms in the shallow domains of orogenic wedges.

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JABAL QUSAYBAH ANTICLINE

This chapter is presented in the form of a manuscript that is currently been prepared for submission to *Tectonics*. The manuscript describes the interactions between left-lateral strike slip, compressional, and extensional deformations during the development of an inversion anticline at the front of the Oman Mountains thrust wedge (Sultanate of Oman). This case study includes (i) structures characterized by multiple reactivations in compression and extensions, and (ii) extensional structures developed as the local consequence of larger-scale compressive structures.

ARTICLE IN PREPARATION

Complex fault-fold interactions during the growth of the Jabal Qusaybah inversion anticline at the western tip of the Salakh Arch, Oman.

Storti F., Balsamo F., Clemenzi L., Mozafari M., Al-Kindy M.H.N., Solum J., Swennen R., Taberner C., and Tueckmantel C.

Article in preparation for submission to *Tectonics*.

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COMPLEX FAULT-FOLD INTERACTIONS DURING THE GROWTH OF THE JABAL QUSAYBAH INVERSION ANTICLINE AT THE WESTERN TIP OF THE SALAKH ARCH, OMAN.

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ABSTRACT

The Jabal Qusaybah anticline is located at the western end of the Salakh Arc, a major salient in the foothills of the Oman Mountains. We performed a structural and petrographical-geochemical study of veins sets and fault zones associated with the development of this anticline. Our data illustrate a complex deformation pattern both in space and time, characterized by the unusual presence of widespread NE-SW left-lateral strike-slip fault zones trending oblique to the E-W fold axial strike, and of abundant and well developed N-S fold-perpendicular extensional fault zones associated with axial bulging and dilation in the central region of the anticlinal crest. Field and reflection seismic data support development of the Jabal Qusaybah anticline by positive inversion of an Early Turonian extensional growth fault zone. We propose an explanation for these complex fault-fold interactions by involving the Early Cambrian Ara evaporites and a possible foredeep-parallel tapering of their sedimentary/tectonic overburden.

4.1 INTRODUCTION

Fold kinematics in shallow thrust-fold belts is driven by a number of factors, including the intimate interplay with underlying faults (e.g., Suppe, 1983; Richard et al., 1991; Fischer et al., 1992; Tavani et al., 2006), the mechanical stratigraphy of the deformed multilayer (Corbett et al., 1987; Woodward & Rutherford Jr, 1989; Protzman & Mitra, 1990; Gross, 1995; Couzens & Wiltchko, 1996; Fischer & Jackson, 1999; Chester, 2003; Di Naccio et al., 2005), the overburden (Chester et al., 1991; Jamison, 1992; Lemiszki et al., 1994), and the ratio between tectonic uplift and syntectonic sedimentation rates (e.g.,

Storti & Salvini, 1996). Fold development in curved deformation front segments may imply further complexities because the formation of salients and recesses is triggered by local variations of the geological conditions at the wedge toe (e.g., Macedo & Marshak, 1999), including the presence of inherited pre-orogenic fault zones. The arcuate shape of salients and recesses is typically accompanied by non-cylindrical deformations and partitioning between dip-slip and strike-slip faulting (e.g., Marshak, 1988). In such a geological framework, the expected folding-related deformation pattern may significantly deviate from the classical longitudinal and transversal distribution of deformation structures (e.g., Stearns, 1968; Storti & Salvini, 1996) described in cylindrical thrust-related folds (e.g., Hancock, 1985; Cooper, 1992; Tavani et al., 2006, 2008). Given the importance that fault-fold growth in thrust-fold belts and accretionary prisms has for both economical and seismic hazard mitigation purposes, descriptions of the kinematic evolution of anticlines developed within thrust salients and recesses can provide effective analogues to support predictions of deformation patterns and related fluid flow in comparable subsurface settings.

In this paper, we describe the type and orientation of faulting- and folding-related deformation structures exposed in the Jabal Qusaybah anticline, which is located at the western inflection point of the Salakh Arch, in northern Oman. Petrographical and isotopical characterization of vein networks provide additional constraints to the relative chronology of folding- and faulting-related deformation structures established in the field. By combining field and laboratory information, an evolutionary pathway is proposed for the Jabal Qusaybah anticline where positive fault inversion, strike-slip and extensional faulting, and fold amplification, coevally interact.

4.2 GEOLOGICAL OVERVIEW

The Jabal Qusaybah anticline is located at the western termination of the Salakh Arch, the large salient that characterizes the southern deformation front of the Northern Oman Mountains (Figure 4.1). The stratigraphic succession in the footwall of the ophiolitic nappes includes Precambrian carbonates and siliciclastic rocks, a Paleozoic section dominated by siliciclastic rocks, and Mesozoic to Cenozoic carbonates with subordinated interlayered siliciclastic formations. The presence of the Early Cambrian Ara evaporites (Fahud and Ghaba salt basins; Figure 4.1) supports the inference that the basal thrust of the Salakh Arch is localized within these evaporites (Hanna, 1990; Al-Kindy, 2006; Filbrandt et al., 2006). The Cretaceous stratigraphic pile starts with the platform carbonates of the Kahmah group, separated by an unconformity from the shales and sandstones of the Nahr Umr Formation; another major unconformity separates the latter from the carbonate Natih Formation whose upper part is in turn eroded and

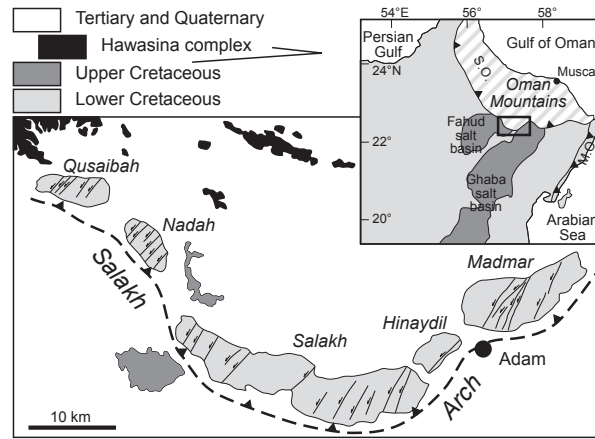


Figure 4.1: Geological sketch map of the Salakh Arc, northern Oman. The inset shows the location of the arch at the southern deformation front of the Oman Mountains (after [Al-Kindy, 2006](#); [Filbrandt et al., 2006](#)). The schematic location and orientation of left-lateral strike-slip fault zones in Jabal Madmar is from [Bazalgette \(2004\)](#). M.O. means Masirah Ophiolite; S.O. means Samail Ophiolite.

unconformably overlain by the shales and sandstones of the Aruma Group (e.g., [Homewood et al., 2008](#), Figure 4.2).

Jabal Qusaybah exposes an about 8 km long, 3 km wide anticline (Figure 4.3) where the Cretaceous Natih Formation and, in particular, the lithostratigraphic members A (Late Cenomanian,-Early Turonian), B (Middle-Late Cenomanian), and C (Middle Cenomanian) crop out. The former mainly consists of platform carbonates and overlies the intrashelf basinal carbonates of the B member. A more competent horizon occurs within Natih B and provides a useful key layer for stratigraphic correlations. Natih C mainly consists of platform carbonates and its top is characterized by a yellowish horizon of coarse-grained rudist debris forming a tidal sandwave (e.g., [Hanna & Smewing, 1996](#); [Homewood et al., 2008](#), Figures 4.2, 4.4) .

The hinge line of the Jabal Qusaybah anticline strikes almost E-W and the exposed cross-sectional fold shape is open and slightly asymmetrical to the north (Figure 4.3a). A characteristic structural feature of this anticline is the presence of linked NE-SW left-lateral strike-slip and N-S extensional fault zones, the latter being concentrated in the central sector of the fold where a peculiar symmetrical central bulge occurs (Figures 4.3, 4.5b). Left-lateral fault zones have an oblique trend with respect to the fold strike and N-S extensional fault zones are confined within strike-slip-bounded compartments that characterize the central part of the fold; otherwise they are located at the southwestern tip of the left-lateral strike-slip fault zones where they terminate within the anticline. As a whole, the fault network creates a left-lateral transtensional horsetail array localized in the longitudinal axial bulge of the anticline (Figure 4.3). A segmented N-S right-lateral strike-slip fault zone occurs to the west and is localized along

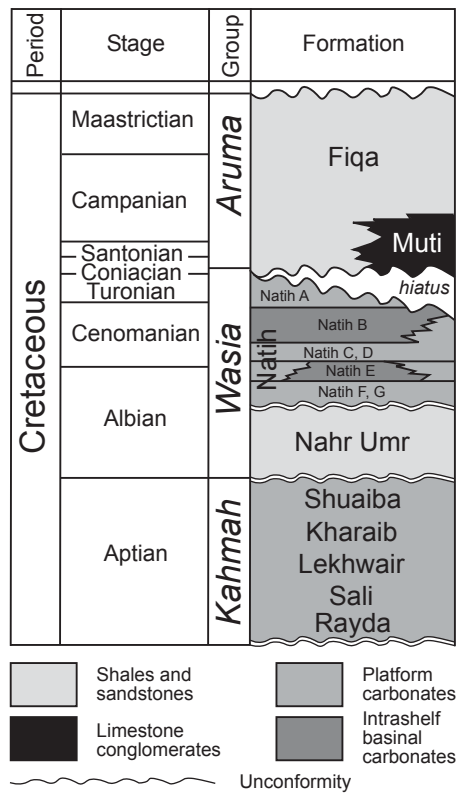


Figure 4.2: Stratigraphic column of the Cretaceous rocks exposed in the Oman Mountains (after Filbrandt et al., 2006; Home-wood et al., 2008).

the hinge between the flat-lying crestal region, and the western limb of the central axial bump (site G in Figure 4.3).

4.3 STRUCTURAL DATA

Deformation structures exposed in the Jabal Qusaybah anticline consist mostly of calcite veins, extensional and strike-slip fault zones, and stylolites. Bedding-parallel stylolites are typically overprinted and overprint a first generation of near-vertical, white to brown E-W, N-S, and NE-SW trending calcite veins (Figure 4.6a), the latter two sets associated with left-lateral strike-slip fault zones. Most calcite veins, however, strike either parallel or perpendicular to bedding regardless of its orientation and location within the anticline (Figure 4.7). In the poorly layered Natih C carbonates, NE-SW striking veins also developed as conjugate shears of the NNE-SSW set (Figure 4.7c, d). In many cases, in the Natih C the latter vein population shows evidence of re-activation as small-scale right-lateral strike-slip faults (Figure 4.7e). Analysis of the entire background vein dataset indicates that many of them strike NNE-SSW, with an additional set striking almost E-W, i.e. perpendicular and parallel to the fold axial strike, respectively (Figure 4.7h). Abutting relations indicate that in many cases veins striking parallel to bedding are confined in-between veins striking in the dip direction of beds (Figure 4.7f). However, the opposite relative chronology also occurs. Tectonic stylolites trending parallel to the fold axial strike occur at some places but are not wide-

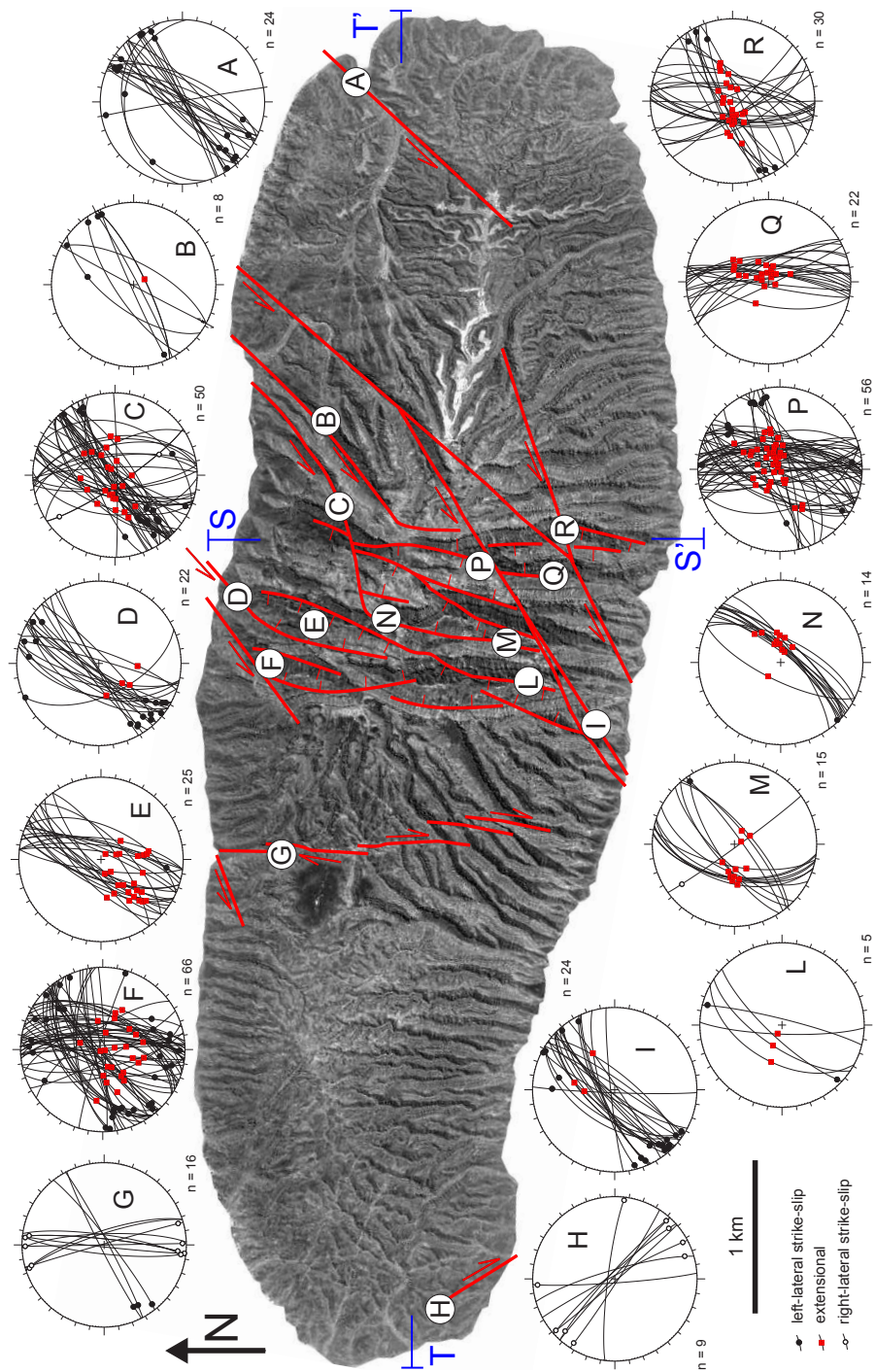


Figure 4.3: Network of major fault zones mapped in Jabal Qusaybah (image from Google Earth) and stereographic projections of fault data collected in field localities A to R. The tip points of cross-sections S-S' and T-T' (Figure 4.5) are indicated. In all figures, stereographic projections of planar surfaces are on the Wulff equal angle net, lower hemisphere; pole contours are made on the Schmidt equal area net, lower hemisphere; when not specified, contouring is at 5% intervals.

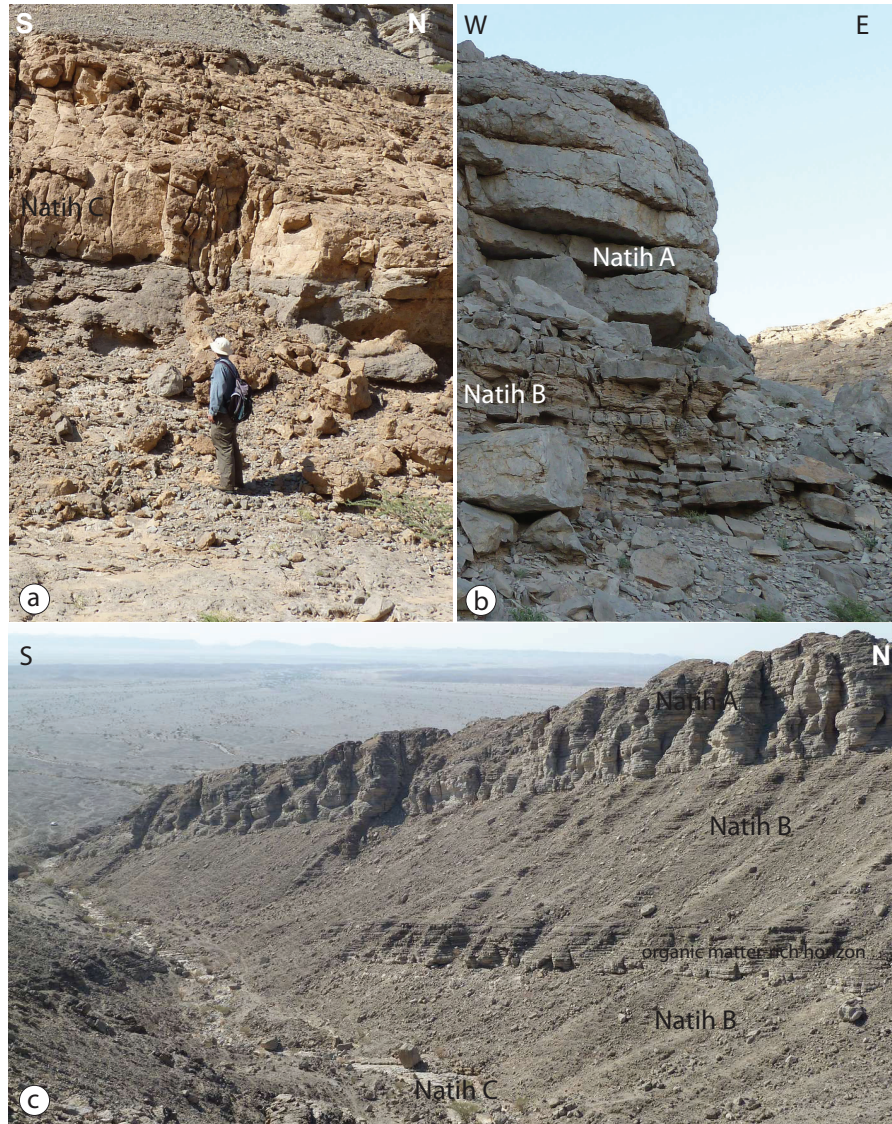


Figure 4.4: a) Yellowish coarse bioclastic grainstone with large cross-beds at the top of Natih C; field locality C in figure 3. b) Detail of the stratigraphic contact between Natih A and B lithostratigraphic members; field locality P in figure 3. c) Panoramic view of the southern limb of the Jabal Qusaybah anticline (photo from locality P in figure 3, looking southward) showing the appearance of the exposed stratigraphic succession from Natih C to A.

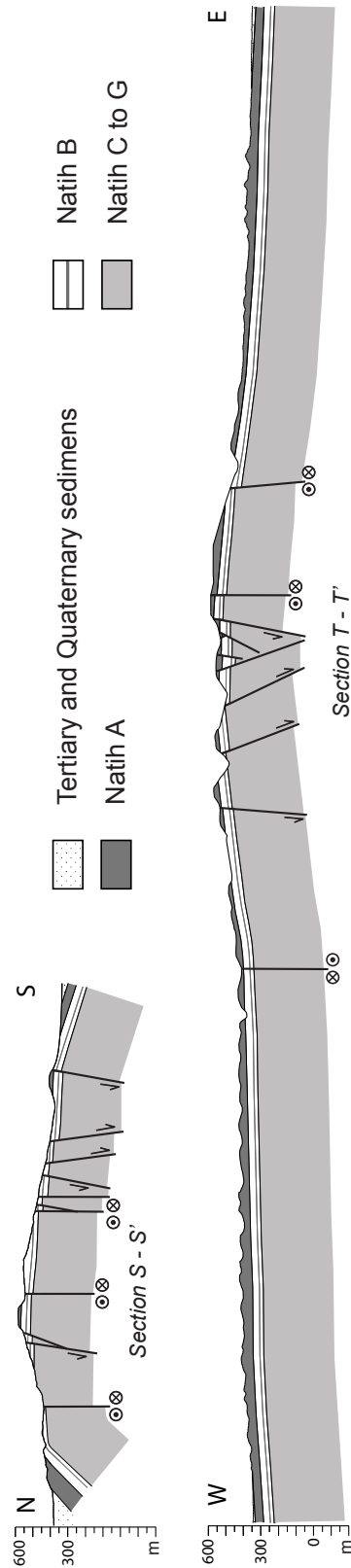


Figure 4.5: Transversal (S-S') and longitudinal (T-T') geological cross sections illustrating the main geometric features and fault zone distribution in the Jabal Qusaybah anticline. See Figure 4.3 for location.

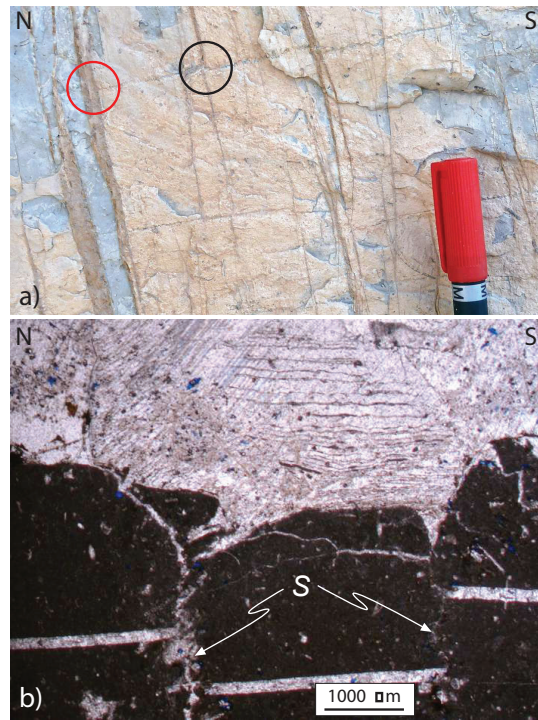


Figure 4.6: a) Cross-sectional view of a thick Natih A limestone bed showing mutually overprinting relations between bedding parallel stylolites and veins (field locality E in Figure 4.3). A stylolite overprints a vein in the area indicated by the black circle and is overprinted by another vein, striking parallel to the previous one, in the area indicated by the red circle. b) Microphotograph from a bedding-parallel thin section (field locality E in Figure 4.3) showing E-W near vertical stylolites (S) overprinting a N-S near vertical thin calcite vein. They are in turn overprinted by a N-S near vertical large calcite vein with evidence of crack-seal opening (plane polarized light).

spread at the outcrop scale. In most cases they cut a first generation of N-S veins and are in turn overprinted by younger veins striking parallel to the older ones and characterized by evidence of multiple opening events (Figure 4.6b).

At some field localities, evidence for positive inversion of pre-existing E-W extensional fault zones was found. A larger scale example is exposed on the eastern cliff of the N-S trending wadi at field locality G (Figure 4.3), where bedding pattern and structural data support the initial occurrence of a complex, northward dipping extensional fault zone affected by subsequent reverse motion (Figure 4.8a-d). Two calcite vein sets are associated with two fault zones exposed at the base of the cliff: the older vein set strikes WNW-ESE and has a near vertical dip; it is overprinted by the younger set with the same strike but a very shallow dip (Figure 4.8c). The first vein set indicates early extensional faulting that, for fault segment 1, is supported by a stratigraphic throw of 4.38 m. On the other hand, the second vein set supports the reversal of fault motion. An extensional throw of 4.26 m

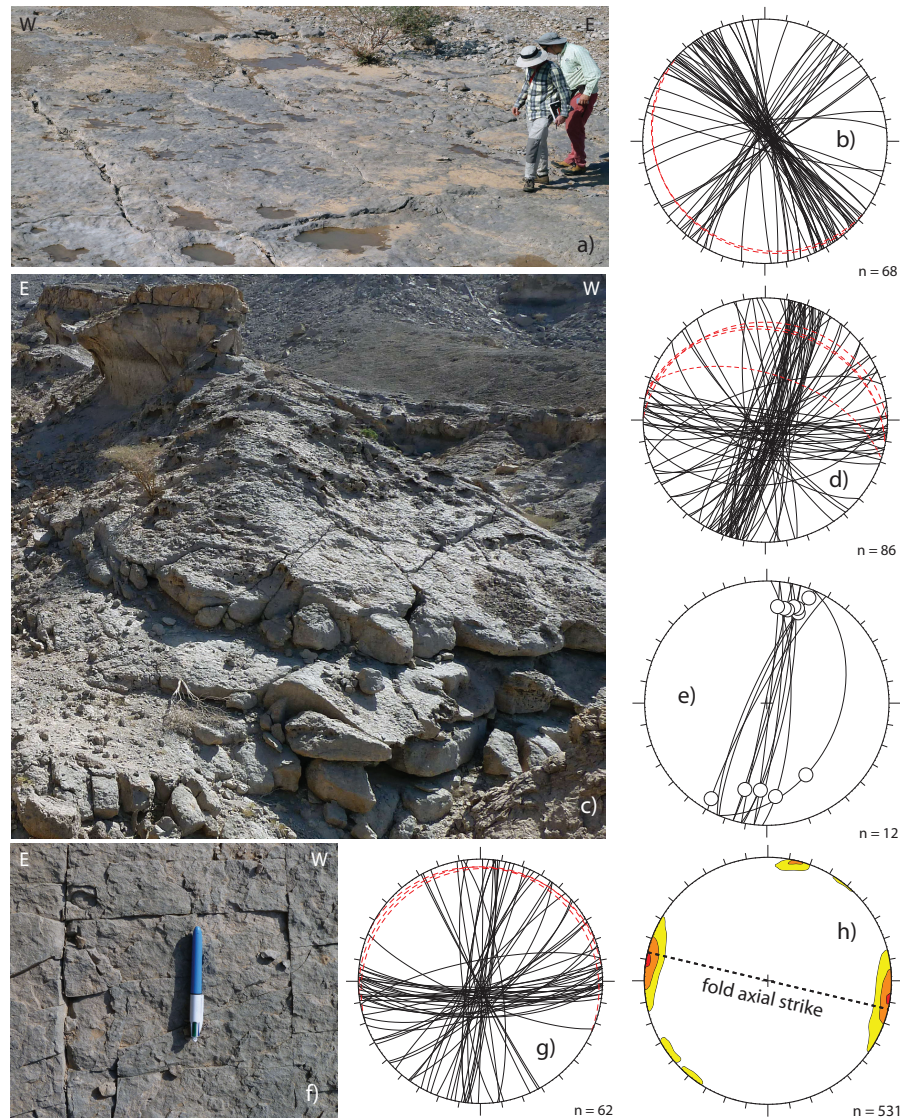


Figure 4.7: a) Pavement view of NW-SE and NE-SW calcite veins and their abutting relations; Natih A, field locality H in Figure 4.3. b) Stereographic projection of vein data collected in the field locality shown in a). c) Panoramic view, looking southward, of Natih C exposed in the forelimb of the Jabal Qusaybah anticline to show the occurrence of different vein sets (field locality E in Figure 4.3). d) Stereographic projection of vein data collected in the field locality shown in c). e) Stereographic projection of NNE-SSW striking veins re-activated as small-scale right-lateral strike-slip faults (data collected in the same field locality). f) Plane view of a Natih B bed affected by two mutually perpendicular vein sets (field locality G in Figure 4.3). g) Stereographic projection of vein data collected in the field locality shown in f). h) Contoured poles to vein data collected at several field localities, far from fault zones.

characterizes fault segment 2, which is characterized by near vertical calcite veins in its damage zone. However, calcite slickenfibers on slip surfaces in the whole outcrop indicate the occurrence of both extensional and contractional kinematics, the latter being the younger motion (Figure 4.8d).

Positive inversion of fault zones exposed at field locality C is indicated by calcite shear vein geometry on slickensides (Figure 4.8e, f). In both cases, the stratigraphic throw is still extensional. In a fault zone to the north of locality C, the former extensional kinematics is indicated by abundant fault-parallel calcite veins, while kinematic indicators on slickensides support reverse slip (Figure 4.8g). In the last two fault zones, positive inversion is also indicated by the presence of a gentle anticline in the fault hangingwall (Figure 4.8f, g). Clear evidence of positive inversion was found at field locality E, where in a ENE-WSW to E-W trending fault zone most subsidiary E-W faults in the damage zone still preserve their extensional kinematics and are associated with fault-parallel, steeply dipping calcite veins (Figure 4.9a-c). However, the thick calcite vein infilling the ENE-WSW master slip surface shows slickenfibers indicating reverse shear. This is supported by microstructural evidence in thin section, showing tectonic stylolites that offset the former master slip surface and overprint the thick calcite vein on it (Figure 4.9d). The orientation of stylolites is compatible with a late stage reverse motion. N-S trending calcite veins occur mainly in the hangingwall of this fault zone, predating the E-W ones (Figure 4.9e). At the northern boundary of this fault zone, an ENE-WSW slip surface preserves near horizontal slickolites indicating left-lateral strike-slip shear, overprinted by steeply-dipping calcite slickenfibers indicating extensional kinematics (Figure 4.9f).

Strike-slip fault damage zones contain abundant veins that are frequently arranged in two sets striking parallel (set 1) and at about 45° from the master slip surfaces (set 2), respectively (Figure 4.10). At field locality G, thick calcite veins adjacent to the master slip surface are folded by fault drag (Figure 4.10b). In the same outcrop, the western wall of the right-lateral strike-slip fault zone exposes the Natih B lithostratigraphic member and is characterized by abundant, closely-spaced stylolites striking mainly NW-SE (Figure 4.10c). The same lithology is involved in the ENE-WSW left-lateral strike-slip fault zone exposed at field locality C, where abundant $\sim$ E-W striking stylolites also occur, mutually overprinting with $\sim$ NE-SW calcite veins (Figure 4.10d, e). A few tens of meters to the NE, the previous fabric is overprinted by fault-parallel stylolites and, in turn, by fault-parallel calcite veins (Figure 4.10f). The relative chronology between different stylolite and vein generations is schematically illustrated in Figure 4.10g. Intersections between different calcite vein sets observed in the damage zone of a NE-SW left-lateral strike-slip fault zone at field locality D show that N-S striking veins and NE-SW ones mutually overprint (Figure 4.11). A NE-SW left-lateral strike-slip fault zone

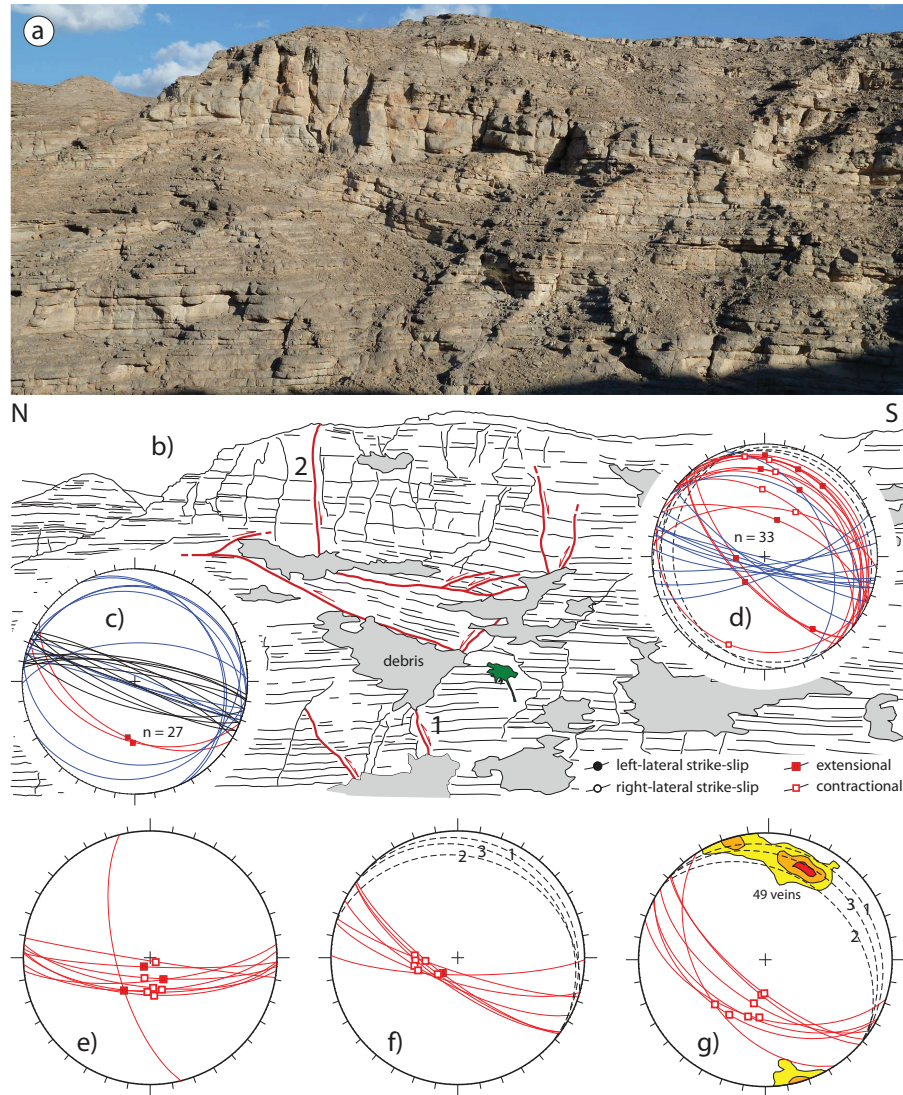


Figure 4.8: a) Panoramic view of the eastern wadi wall at field locality G. Natih A limestone is exposed, showing a bedding and fault pattern that support positive inversion of a pre-existing E-W extensional fault zone. b) Line drawing of the same image. c) Stereographic projection of structural data collected on the two faults exposed in the lower part of the outcrop: faults are in red, veins in blue, and tectonic stylolites in black. d) Stereographic projection of structural data collected in upper part of the outcrop: broken circles indicate the attitude of bedding. e) Stereographic projection of fault data collected westward of field locality C. f) Stereographic projection of fault data collected at field locality C. g) Stereographic projection of fault and bedding data (great circles), and contoured poles to veins collected northward of field locality C. In (f) and (g), 1 refers to the attitude of bedding in the footwall of the inverted fault zone, 2 refers to bedding in the hanging wall close to the fault zone itself, and 3 refers to the average attitude of bedding outside the slightly folded area. In both cases, bedding forms a gentle anticline within the northward-dipping forelimb of the host fold.

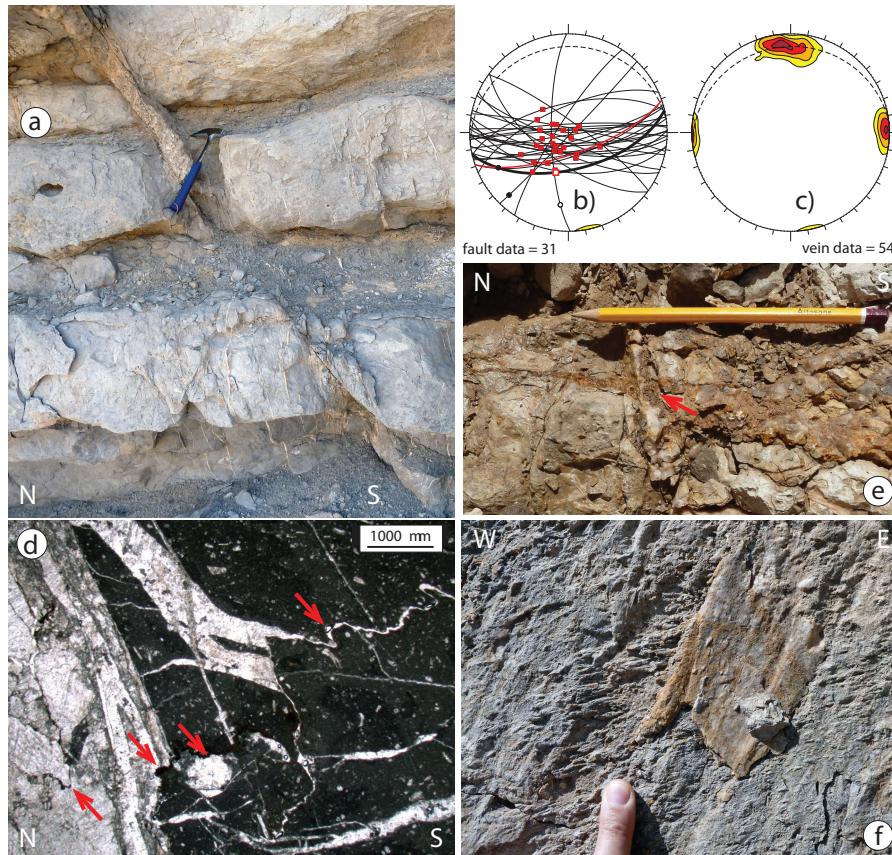


Figure 4.9: Example of positive inversion of an E-W extensional fault zone at field locality E. a) Cross-sectional view of a segment of the fault zone showing abundant calcite veins both in the hangingwall and in the footwall, and the presence of a thick vein infilling the dilated master slip surface (at location of hammer). b) Stereographic projection of fault data collected at this field site; the thicker great circle provides the orientation of the inverted master slip surface. c) Stereographic projection showing contoured poles to veins. d) Photomicrograph showing a detail of the thick vein infilling the dilated master slip surface and the host rock in the hangingwall. Multiple events of calcite crystallization are evident in the vein, with larger twinned crystals occurring in the central sector. The master slip surface is offset by a stylolite that affects also the host rock (arrows 1 and 2). Other dissolution surfaces are evident within both the large vein (arrow 3) and the host rock (arrow 4). In both cases, smaller calcite veins coeval with stylolites formed at some stylolitic teeth (arrows 1 and 5). The geometric relations between the master slip surface and the stylolites and related veins indicate a reverse motion overprinting deposition of the thick calcite infill. e) Plane view of an E-W calcite vein cross-cutting a N-S one. The red arrow indicates the intersection point. f) Detail of an E-W master slip surface showing near horizontal slickenites indicating left-lateral strike-slip motion, overprinted by calcite shear fibers developed during subsequent extensional motion.

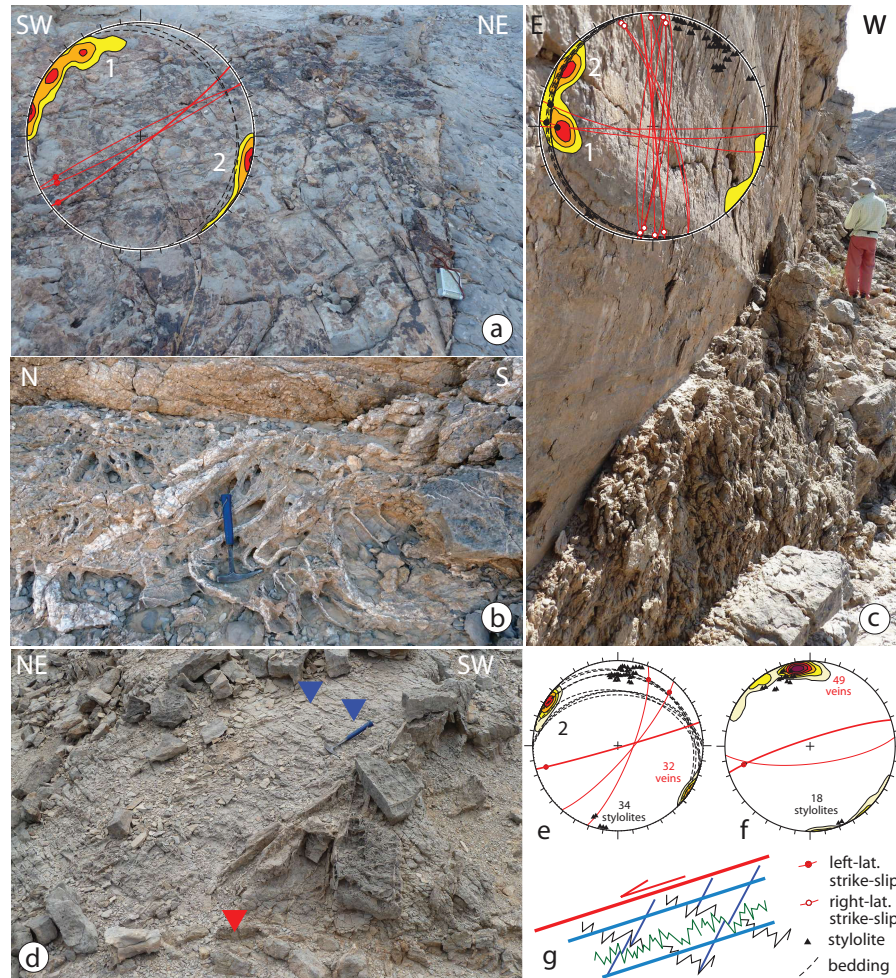


Figure 4.10: Examples of deformation structures in strike-slip damage zones. a) Field locality G: plane view of a NE-SW left-lateral strike-slip fault zone developed in Natih A; note the curved trajectory of calcite veins in the southern damage zone, from fault parallel adjacent to the master slip surface, to almost N-S moving away from it. The stereonet shows structural data collected in the same fault zone: red great circles indicate the geometry and kinematics of fault zones (the trace of the master slip surface is thicker); broken great circles indicate the attitude of bedding; contouring shows the geometry of calcite veins (45 data): maximum 1 refers to fault-parallel veins while maximum 2 indicates veins at about 45° counterclockwise from the master slip surface. b) Plane view of tightly folded calcite veins adjacent to the master slip surface of the main N-S right-lateral strike-slip fault zone exposed at field locality G. c) Cross-sectional view of the corrugated N-S striking master slip surface and closely-spaced NW-SE striking stylolites (foreground) in the western damage zone of the right-lateral strike-slip fault zone exposed at field site G. The master slip surface developed in Natih A, while the western damage zone developed in Natih B. The stereographic projection shows structural data collected in the same fault zone: fault geometry and kinematics is indicated by red great circles; black triangles indicate the attitude of tectonic stylolites (the trace of the master slip surface is thicker); contouring show the attitude of calcite veins (92 data). (Continues on the following page).

Figure 4.10 (*continued*): Note that the N-S fault traces parallel the strike of bedding and of vein set 1, whereas vein set 2 lies at about 45° clockwise from the master slip surface. d) Plane view of the southern damage zone of the NE-SW striking left-lateral strike-slip fault zone exposed at field locality C, Natih B; the red triangles outline the trace of the slip surface, and the blue triangles outline the trend of tectonic stylolites. e) Structural data collected in the same site as shown in d) and contouring of poles to 36 veins (contouring at 12%). f) Structural data collected some tens of meters to the NE along the fault zone; symbols are the same as in (e). g) Sketch map showing the overprinting relations between different vein and stylolite generations, which demonstrate the transition from simple shear, to local fault-perpendicular compression, and then to local fault-perpendicular extension.

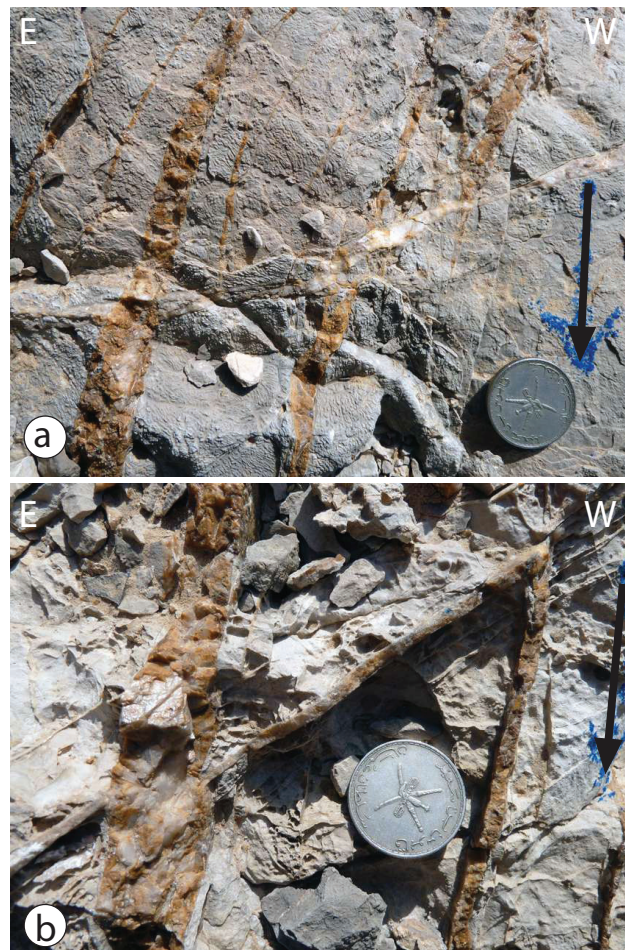


Figure 4.11: Plane views of different intersecting calcite vein generations along a NE-SW left-lateral strike-slip fault zone exposed at field locality D. a) NE-SW striking white calcite vein offsetting a N-S brown calcite vein (blue circle) and in turn offset by another N-S brown calcite vein (red circle). b) NE-SW striking white calcite vein offsetting a N-S brown calcite (blue circle) and in turn offset by another N-S brown calcite vein (red circle).

exposed at field locality A provides relations among different calcite vein generations far from the intensely deformed central axial bulge (Figure 4.12). The oldest calcite veins trend about N-S, at about 60° counterclockwise from the master slip surface (set 1 in Figure 4.12c). A NNE-SSW striking splay fault zone shows the same angular relations with vein set 1 (Figure 4.12d, e). Vein set 1 is overprinted by vein set 2 that trends NNE-SSW, at about 30° ccw from the master slip surface (Figure 4.12f). Finally, vein set 3 cuts across the previous deformation structures and trends about NW-SE (Figure 4.12g).

Extensional fault zones trend mostly N-S to NNE-SSW (Fig. 3). Most of them have an “Andersonian dip”, but near vertical fault zones showing extensional downthrows also occur (e.g. localities C, P, Q, in Figure 4.3). In some cases, extension-related deformation structures are overprinted by right-lateral strike-slip ones (Figure 4.13a, b), while in other cases they are overprinted by left-lateral strike-slip motions, particularly when the former extensional fault zones trend NNE-SSW and in the proximity with the NE-SW left-lateral strike-slip master fault zones (Figure 4.13c). In few cases, such as locality E in figure 3, damage zones preserve the evidence of previous right-lateral strike-slip activity indicated by near vertical stylolites making an acute angle with the strike of the extensional master slip surface (Figure 4.14). In extensional fault damage zones, near vertical thick veins striking parallel to the master slip surfaces are abundant and show clear evidence of multiple calcite infilling events, being the youngest one commonly characterized by euhedral crystals (Figure 4.13d, e). At field locality P, bedding parallel veins are widespread in the footwall of a master extensional fault zone. They include thick wedge-shaped ones that taper away from the fault core for some meters (Figure 4.13f).

The internal structure of most fault cores, both in extensional and strike-slip fault zones, is quite unusual. Typically, they show impressive evidence of dilation and multiple events of calcite precipitation, similar to the internal structure of veins in damage zones (Figure 4.13g). In this case, however, euhedral crystals frequently have very large sizes (greater than 20 cm). In more detail, some fault cores still preserve layers of strongly cemented grey monogenic cataclasite associated with smoothly polished master slip surfaces (Figure 4.15a, b). In places, these cataclastic bodies are silicified. In most cases, they are bounded in one or both sides by layers of yellow-to-reddish cataclasite consisting of limestone clasts embedded into very abundant cement made of large calcite crystals with blocky texture and frequent dark to black hydrocarbon inclusions (Figure 4.15c). This cataclasite generation is in turn frequently affected by further cataclasis (Figure 4.15d), followed by dilation and deposition of multiple generations of layered calcite growth bands and large calcite crystals and aggregates (Figure 4.13g).

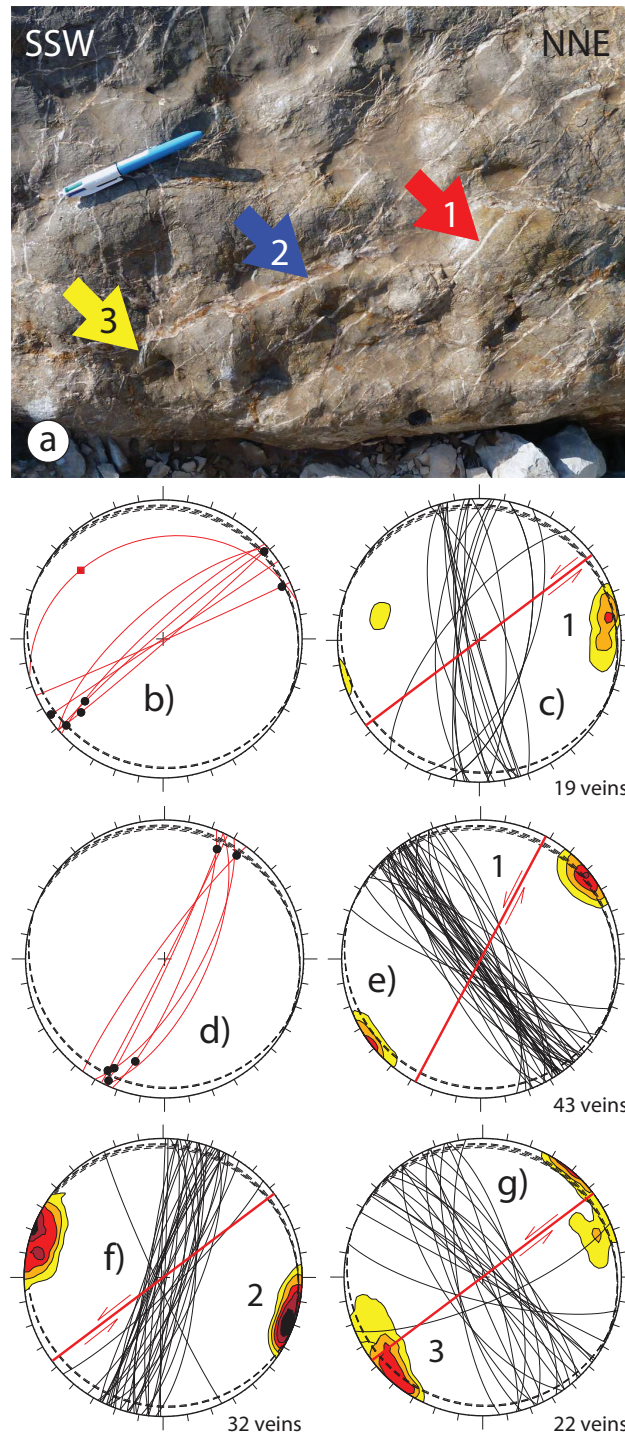


Figure 4.12: Vein network in the damage zone of a NE-SW left-lateral strike-slip fault zone exposed at field locality A. a) Detailed plane view of a damage zone sector showing the presence of three different veins sets indicated by 1, 2, 3. b) Stereographic projection showing fault geometry and kinematics at site shown in a). c) Attitude of set 1 veins with respect to the average trend of the fault zone (shown in red). d) Stereographic projection showing fault geometry and kinematics of an adjacent NNE-SSW splay fault. e) Attitude of set 1 veins with respect to the average trend of the fault zone. f) Attitude of set 2 veins with respect to the average trend of the fault zone at site shown in a). g) Attitude of set 3 veins with respect to the average trend of the fault zone at site shown in a).

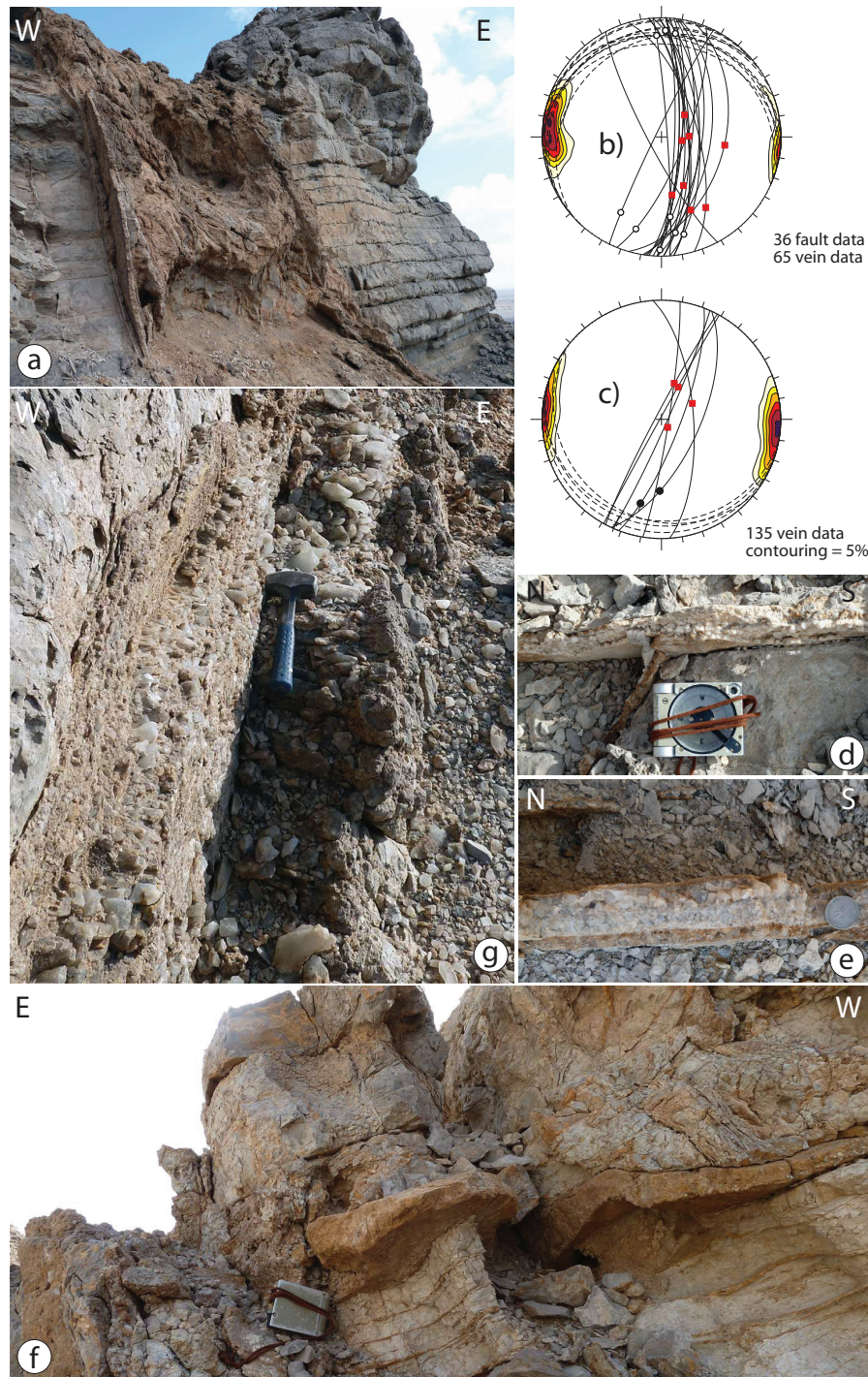


Figure 4.13: a) Cross-sectional view of an extensional fault zone exposed at field locality F; the brownish-orange color highlights abundant iron-rich calcite precipitation. b) Structural data collected at this field locality. c) Structural data collected along an extensional fault zone exposed at field locality L. d) Plane view of a N-S striking white calcite vein showing evidence of multiple infilling events; field locality C. e) Plane view of a N-S striking brown to white calcite vein showing evidence of multiple infilling events; field locality P. f) Cross sectional view of bedding-parallel veins in the footwall damage zone of the N-S master extensional fault zone exposed at field locality P. g) Plane view of a N-S extensional fault core showing evidence of multiple infilling events by alternating layers of large euhedral calcite crystals and aggregates; field locality L. Note that kinematic indicators on stereonets are as given on Figure 4.3.



Figure 4.14: a) Cross-sectional view of a N-S extensional fault zone; the stereonet in the inset shows the accompanying structural data. b) detailed plane view of a shear lens preserved along the master extensional slip surface: closely-spaced, near vertical N-S stylolites are evident.

4.4 PETROGRAPHIC AND GEOCHEMICAL DATA

The different vein sets, both in the background deformation pattern, in fault damage zones, and within fault cores, were systematically sampled for petrographic and geochemical analysis. Based on petrography, geochemistry and structural features, veins can be subdivided into three main groups, each representing unique precipitation condition attributed to fluid circulation linked to different stages of tectonic deformation (Figure 4.16). Here representative samples for relative chronology purposes are described; the complete dataset will be presented in a specific paper.

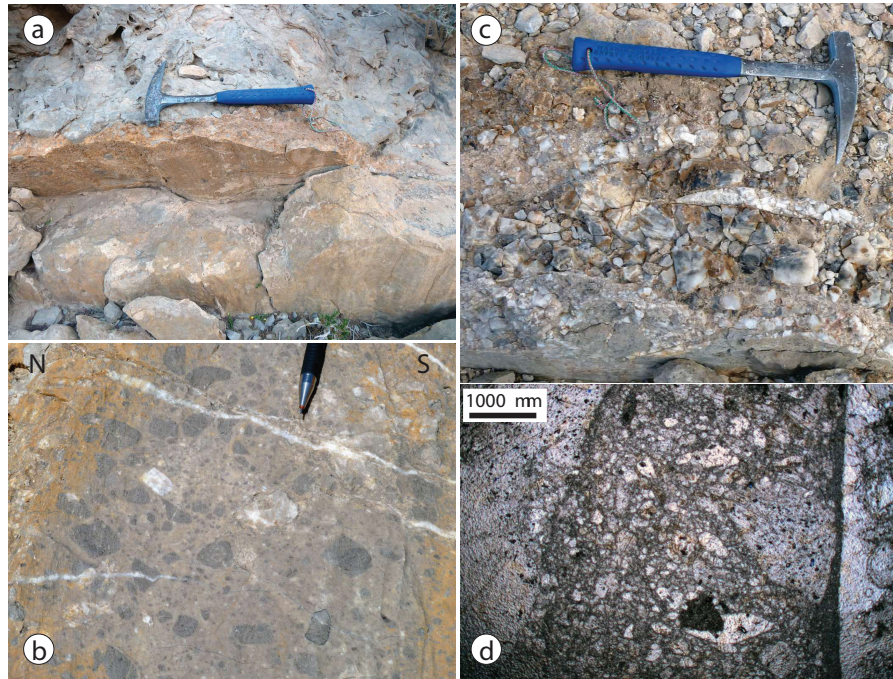


Figure 4.15: Main features of fault cores. a) Layer of strongly cemented monogenic cataclasite (1, in the foreground) that underwent further dilation on one side and infilling by iron-rich calcite (2); field locality C. b) detail of a polished master slip surface in strongly cemented monogenic cataclasite: the presence of white calcite veins cutting through the cataclasite and striking perpendicular to abrasion-related striations indicate fault activity after cataclasite cementation; field locality P. c) Detailed plane view of calcite crystals and aggregates deposited in the hanging-wall side of a dilated master slip surface at field locality C; several events of cracking and sealing by different calcite generations are evident. d) Photomicrograph of a calcite vein with hydrocarbon inclusions (some of them are highlighted by red circles), affected by further cataclasis; same fault zone as in c).

4.4.1 Methods

To evaluate the iron content in dolomite and calcite, polished rock slabs and thin sections were stained using Alizarine Red S and potassium ferricyanide (Dickson, 1966). Cold cathodoluminescence microscopy was carried out using a modified Technosyn cathodoluminescence device model 8200, Mark II, operating at 10 - 16 kV gun potential, 200–400 μ A gun current, 0.05 Torr vacuum and 5 mm beam width. Samples for stable oxygen and carbon isotopes analysis were collected employing a dental drill and Merkantec micromilling device analysed at the Friedrich - Alexander Universität (Erlangen - Nürnberg, Germany), and the Vrije University Amsterdam (The Netherlands) using a ThermoFinnigan Five Plus and Finnigan DeltaPlus IRMS mass-spectrometer, respectively. All values are reported in per mil relative to V-PDB.

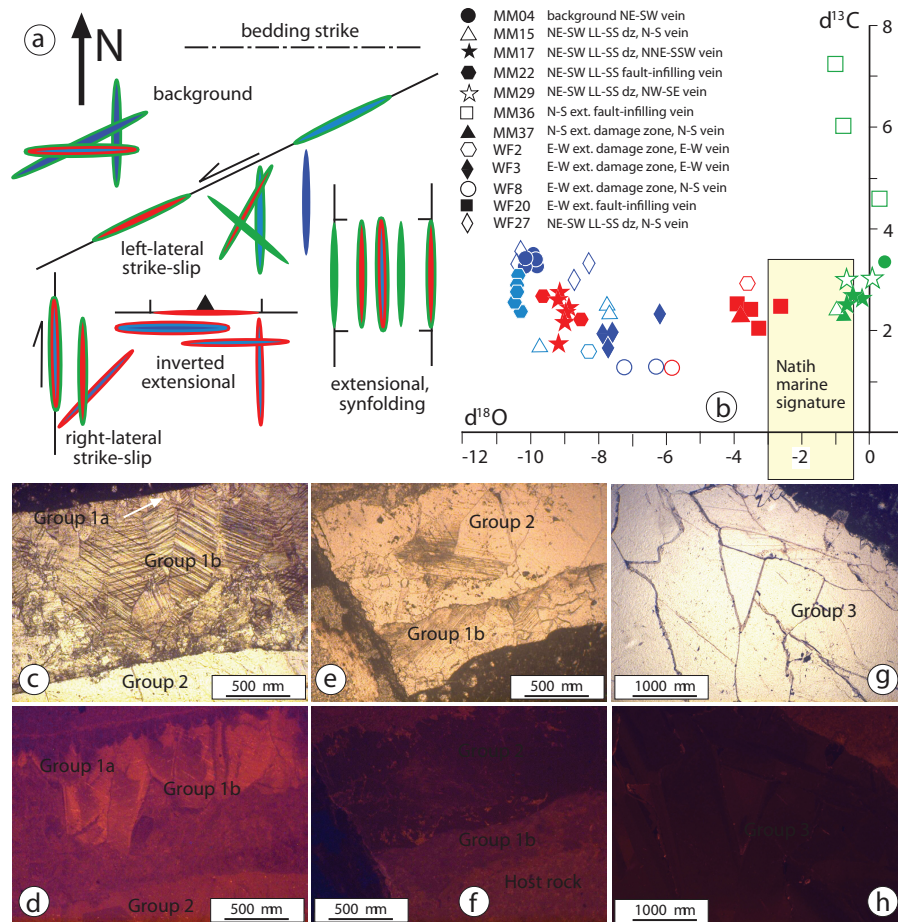


Figure 4.16: Petrographical and geochemical characterization of calcite vein sets. a) Conceptual sketch map of fault-vein arrays in Jabal Qusaybah: veins are oriented with respect to the geographic north and their colours indicate the relative chronology of calcite precipitation (multiple colours indicate vein re-opening): dark blue indicates the Group 1 oldest veins; pale blue indicates the younger veins belonging to the same group; red indicates Group 2 veins, and green indicates Group 3 veins. b) Cross-plot $\delta^{13}\text{C}$ versus $\delta^{18}\text{O}$ of the analyzed veins and fault core infillings; same colour code as in a). c) Sample WF03: photomicrograph in plane polarized light of Group 1 calcite generation, re-opened by Fe-rich Group 2 calcite. Note the deformational twins in Group 1 crystals. d) Cathodoluminescence image of the same view as shown in c). Note the two different CL patterns in Group 1 calcites (1a and 1b, respectively). The non-CL crystals predate the crystals with brown CL possessing a slight zonation pattern. e) Sample WF02: photomicrograph in plane polarized light of Group 1b calcite re-opened by Fe-rich Group 2 calcite. Note the deformational twins in Group 1 crystals. A fragment of Group 1 vein is embedded in Group 2 blocky crystals (untwinned). f) Cathodoluminescence image of the same view as shown in e), showing brown CL pattern in twinned Group 1 calcite and dull brown in Group 2 calcite. g) Photomicrograph in plane polarized light of Group 3 calcite crystals. h) Cathodoluminescence image of the same view as shown in g), illustrating brown CL pattern with sector zonation.

4.4.2 *Vein Group 1*

This group includes the first generations of veins formed in Jabal Qusaybah. Far away from fault zones, they occur as isolated features typically striking parallel to bedding dip, E-W, and NE-SW. Within fault damage zones, these veins strike N-S along NE-SW left-lateral strike slip faults, NE-SW along N-S right-lateral strike-slip faults, and E-W along E-W extensional faults (Figure 4.16a). Group 1 veins consist of non ferroan, cloudy white calcite crystals with a thickness not exceeding 1-2 cm. They are intensely affected by tightly spaced mechanical twinning (Types II and III of Burkhard, 1993). The cathodoluminescence pattern is characterized by non-luminescence, brown and dull brown exhibiting slight zonation (Figure 4.16c, d). Based on abutting relations, the non-CL crystals predate the crystals with brown CL. Accordingly, they can be divided into two sub-groups having similar geochemical properties. The younger sub-group contains the older calcite infills along NE-SW left-lateral strike-slip and N-S right-lateral strike-slip fault cores (Figure 4.16a). The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ composition of Group 1 veins varies between 10.4 and 6.0‰ V-PDB and between +1.3 to +3.5‰ V-PDB, respectively (Figure 4.16b). For comparison purposes, the stable isotope signatures of the Natih Formation marine carbonates measured in Jabal Qusaybah range for $\delta^{18}\text{O}$ between -3.0 and -0.5‰ V-PDB and for $\delta^{13}\text{C}$ between -1.0 and +3.5‰ V-PDB (Immenhauser et al., 2000).

4.4.3 *Vein Group 2*

Group 2 veins are the most abundant in the Jabal Qusaybah anticline. They typically occur as new calcite infills precipitated along re-opened E-W, NE-SW, and N-S Group 1 veins, and also as newly formed veins in E-W extensional and NE-SW and N-S strike-slip fault damage zones and fault core infills (Figure 4.16a). However, they are particularly abundant and well developed along N-S striking extensional fault zones located in the central sector of the Jabal Qusaybah anticline. Group 2 veins contain slightly ferroan to ferroan randomly oriented blocky calcites. The early crystals are characterized by a milky white to pale yellow colour and intense deformational twins (Types II and III of Burkhard, 1993) with undulose extinction indicating tectonic deformation. On the other hand, late crystals in both fault core infills and veins have black, grey or translucent white colours, depending on amount of crystal impurities, with pervasive brown alterations in hand specimen. In thin section they have less or no twinning (Figure 4.16e). The CL patterns of Group 2 calcites are relatively homogenous brown in non-altered samples. However, also blotchy outlines occur which are interpreted to correspond with recrystallised portions (Figure 4.16f). The $\delta^{18}\text{O}$ signatures of this group of calcite display a wide range vary between -9.6 and 2.6‰ V-PDB. This large

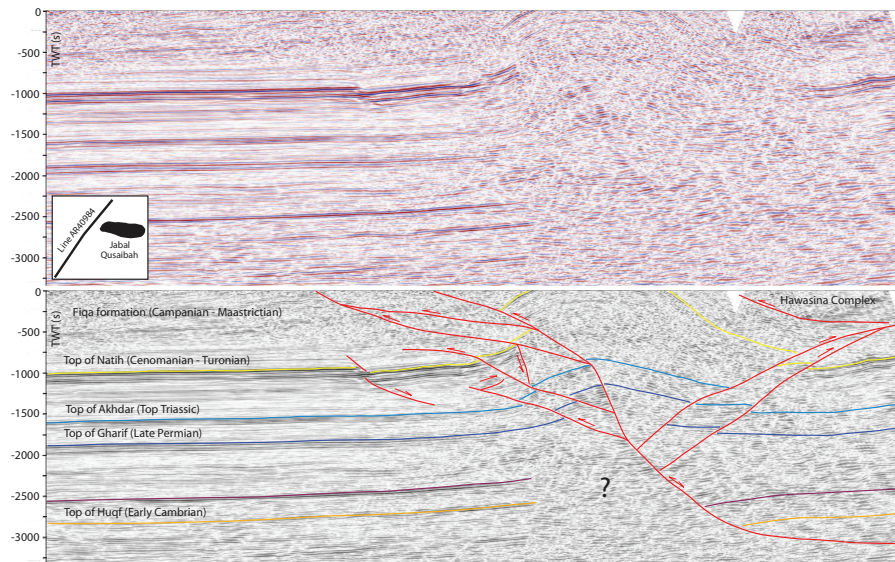


Figure 4.17: Line AR40984 and its proposed interpretation, illuminating the cross-sectional geometry in the western pericline of the Jabal Qusaybah anticline. See text for details.

spread likely reflects the existence of several sub-generations of calcite crystals that, for the purposes of this work, are grouped together. On the other hand, $\delta^{13}\text{C}$ composition displays a relatively narrow range of +1.3 to +2.7‰ V-PDB (Figure 4.16b).

4.4.4 Vein Group 3

Group 3 veins include the last calcite generations that cemented fractures in the Jabal Qusaybah anticline. They are best developed along the N-S extensional fault zones located in the central axial bulge of the crestal region, but they also overprint the older structural fabric (Figure 4.16a). Group 3 veins size spans from mm-scale microfractures to meter-scale fault infills. They possess a variable low to intermediate Fe content and their typical colour is white to translucent white, with grey growth zones characterized by minor yellow to brown alterations if compared to group 2 calcites. The crystals are commonly elongate with axial length perpendicular to the fracture walls, exhibiting brown CL colour with sector zonation (Figure 4.16g, h). The size of crystals reaches up to about 30 cm in fault infillings characterized by bladed to scalenohedral crystals. The $\delta^{18}\text{O}$ compositions demonstrate relatively less depleted values ranging between -1 and +0.1‰ V-PDB. The $\delta^{13}\text{C}$ composition varies between +3.0 and +7.3‰ V-PDB (Figure 4.16b).

4.5 SEISMIC REFLECTION DATA

The two-dimensional seismic profile AR40984 (Al-Lazki et al., 2002) was re-interpreted to evaluate the subsurface structure of the Jabal Qusaybah anticline. The profile was acquired to the west of the anticline, along a NE-SW trajectory (Figure 4.17). Faults and the formation tops of Natih (Cenomanian-Turonian), Akhdar (top Triassic), Gharif (Late Permian), Huqf (Early Cambrian), and the base of the allochthonous Hawasina Complex, were interpreted across the seismic section. The most evident tectonic feature is an antiformal structure that occurs in the northeastern half of the profile. The sharp interruption of the seismic signals that occurs in the core of the antiform has been interpreted as the evidence of a pre-existing extensional fault zone that then underwent positive inversion, as indicated by: (i) the harpoon structure imaged in the folded seismic interval between top Akhdar and top Natih, with quite evident growth wedge geometries illuminated in the upper part of the latter formation; (ii) contractional displacement along the master fault zone is larger for top Natih and then decreases in the older reference layers; (iii) the interpreted position of top Huqf in the fault hangingwall, as well as that of the overlying layer, still maintain an extensional displacement (Figure 4.17). Both footwall shortcut faults and backthrusts splay off the inverted master fault zone. A northward-dipping extensional fault zone is preserved in the footwall of the master fault zone, as well as an originally south-dipping folded one. In the frontal part of the anticline, the Late Cretaceous Fiqa sediments have a gentle southward dip and onlap onto the folded forelimb strata. The area of increased noise that vertically projects below the inversion anticline does not allow detecting with some confidence the geometry of the interpreted reference layers in this part of the seismic profile.

4.6 DISCUSSION

4.6.1 *Interpreted deformation pattern evolution*

The combination of structural, petrographical, and geochemical data allows us to propose the following evolutionary steps for the deformation pattern of the Jabal Qusaybah Anticline (Figures 4.18, 4.19). For this purpose, vein geometry has been referred to an E-W bedding strike in order to remove possible bias created by non-cylindrical folding. This is the case of NW-SE veins, which we interpret as syn- rather than pre-folding (e.g., Bazalgette, 2004) deformation structures developed in NW-SE trending fold segments (see Figure 4.7) and, hence, drawn as E-W features in our evolutionary sketches.

The spatial distribution of the oldest Group 1 veins indicates that extension along E-W fault zones occurred concurrently with burial-related compaction and dissolution. The presence of same generation

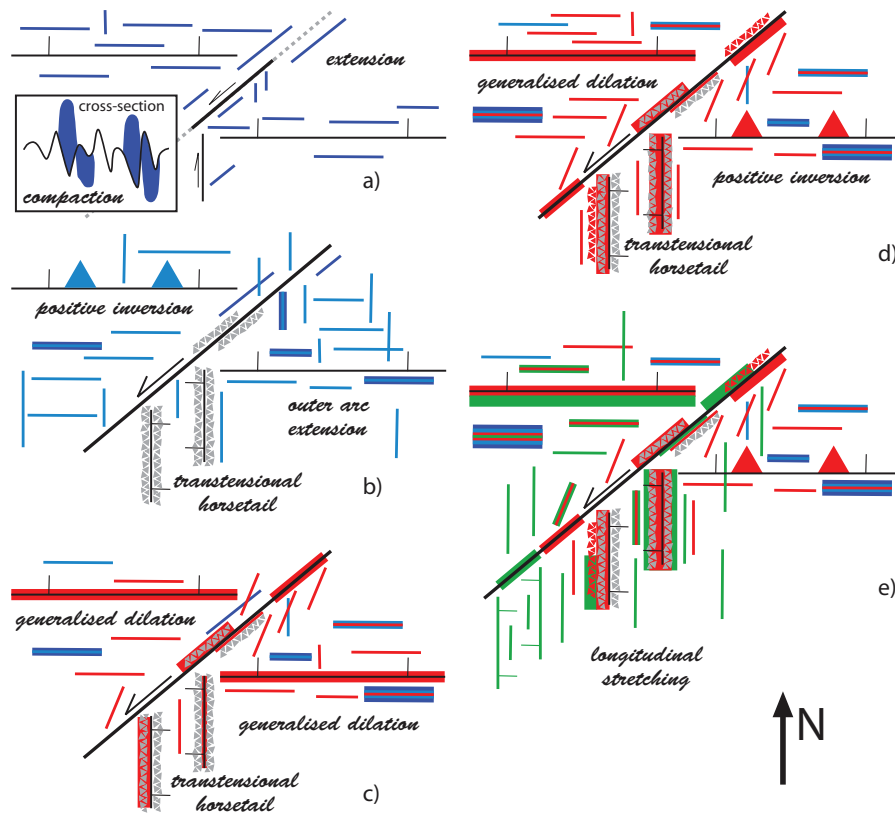


Figure 4.18: Conceptual sketch maps showing the proposed evolution of the deformation pattern in the Jabal Qusaybah anticline. Same colour code as in Figure 4.16. See text for details.

NE-SW veins in the damage zones of NE-SW left-lateral strike-slip fault zones can be explained by re-activation of inherited weakness zones at depth (e.g., [Filbrandt et al., 2006](#), Figure 4.19a). Formation of such tensile veins supports the overall extensional stress field triggering this re-activation. Early right-lateral strike-slip motions along N-S fault zones occurred at this stage, possibly induced by tear faulting and favored by the NE-SW left-lateral shearing (Figure 4.18a). The pattern of younger Group 1 veins supports the contrasting evidence of positive inversion along some E-W fault zones (e.g. site G in Figure 4.3), coeval with continuing extension along other ones (e.g., site E in Figure 4.3). Development of N-S extensional fault zones and related early cataclasites in between NE-SW sinistral ones likely occurred at this stage. The presence of similar cataclastic rocks along the latter indicates that NE-SW left-lateral strike-slip faulting possibly triggered formation of the N-S extensional ones in transtensional horsetail arrays accommodating horizontal displacement at fault tips (Figure 4.18b). Reconciling different fault kinematics is possible by assuming a reversal of the regional stress field from extension to contraction. In the early growth stages of the Jabal Qusaybah anticline, pre-existing weakness zones underwent selective re-activation: the weakest E-W extensional fault zones were re-activated as reverse

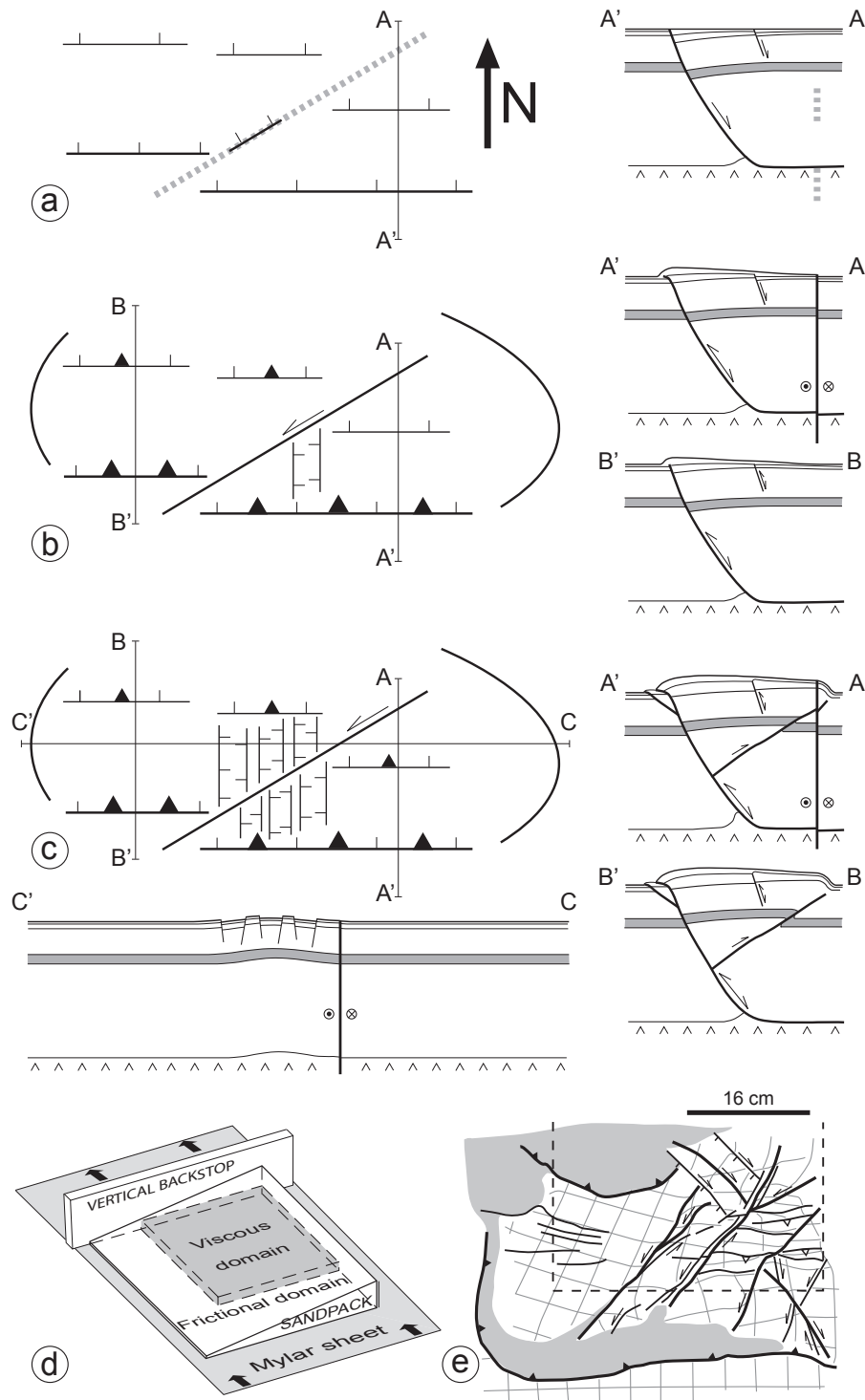


Figure 4.19: a-c), Cartoon showing the proposed evolution of faulting and folding in map view (left) and cross-sections (right) of the Jabal Qusaybah anticline. d) Simplified sketch of the pre-deformational configuration of sandbox experiments described in the text. Note that the sandpack overlying the basal silicone layer (viscous domain) has a gentle taper perpendicular to the plate convergence direction. e) Map view of the tectonic architecture produced at the end of an experiment. d) and e) are simplified drawings after [Storti et al. \(2007\)](#). See text for details.

ones while the stronger fault zones did not fail when compressional deformation started and resumed as tensile deformation structures by outer arc extension during initial fold amplification (Figure 4.19b). The bulk of Group 1 N-S veins developed at this stage by layer-parallel compression, associated with E-W tectonic stylolites and conjugate shear fracturing in Natih C. Formation of veins paralleling the fold axial strike started almost contemporary by outer arc extension, as indicated by the evidence that their abutting relations with veins paralleling the bedding dip direction predated calcite mineralization. Re-activation of NE-SW trending weakness zones as left-lateral strike-slip shear zones fits very well into a compressional environment with a near horizontal N-S trending maximum principal axis of the stress ellipsoid. The evidence that longitudinal (E-W) veins are not common in NE-SW left-lateral strike-slip damage zones, where they are frequently replaced by tectonic stylolites, indicates that folding and sinistral shearing coevally occurred. This is supported by exploitation of some NE-SW fault zones in the crestal area as fold hinge segments. Such a concurrent shearing and folding caused the complex deformation pattern that frequently occurs along the NE-SW left-lateral strike-slip fault zones (e.g., Figure 4.10g).

Formation of Group 2 vein and fault core calcite infills along all pre-existing structural trends supports a period of significant generalized dilation with progressing fold amplification (Figure 4.18c). The abundance and magnitude of N-S dilational features regardless of the position within the exposed Natih formation, indicate that this stage of fold evolution was dominated by stretching parallel to the fold axial strike (Figure 4.19c) with a negligible role of the mechanical stratigraphy to promote selective dilational faulting (e.g., Ferrill & Morris, 2003). The presence of bedding parallel Group 2 thick veins (Figure 4.13f) may support transient fluid overpressuring at this stage. Activity of the NE-SW left-lateral strike-slip fault zones and related N-S extensional ones continued at this stage, as indicated by cataclasis of earlier Group 2 fault infill calcite aggregates and by formation of younger NE-SW calcite veins overprinting the earlier fabric in left-lateral strike-slip damage zones. Positive inversion of some E-W extensional fault zones also occurred (e.g., site E in Figure 4.3; Figure 4.18d). Deposition of Group 3 calcite infills continued in an overall dilational environment. Such late-stage longitudinal extension can be associated with the final development of the central axial bulge that characterizes the Jabal Qusaybah anticline (Figures 4.5, 4.18e). Activity of the NE-SW left-lateral strike-slip fault zones still occurred at this stage, despite being subordinate to longitudinal bulging and stretching.

4.6.2 *Framing the Jabal Qusaybah anticline into the regional picture*

Both field and subsurface data indicate that the Jabal Qusaybah anticline formed by positive inversion of a pre-existing extensional fault zone that developed during deposition of Natih A sediments, possibly in Early Turonian times. Evidence for positive fault inversion is common in the foreland of the Oman Mountains, mainly by reactivation of E-W to WNW-ESE extensional fault zones formed in Late Cretaceous times, during the obduction of oceanic lithosphere onto Mesozoic platform carbonates of the continental margin in North Oman (Robertson, 1987; Loosveld et al., 1996; Filbrandt et al., 2006). Specifically, positive fault inversion in the Jabal Qusaybah anticline occurred either in Santonian (Al-Lazki et al., 2002) or Campanian times (Hanna & Smewing, 1996), depending on whether the Fiqa sediments are interpreted as deposited after or during fold growth, respectively. Subsurface data in Figure 4.17 do not allow us to unequivocally support one of the two hypotheses because Fiqa sediments do not show clear growth wedge geometry (e.g., Hardy & Poblet, 1995) and their gentle tilt and deformation could result from post-depositional folding as well.

Depending on whether extensional faulting occurred during deposition of the Natih carbonates or the overlying unconformable pelagic sediments of the Muti and lower Fiqa formations, this deformation can be associated with different stages of the expected evolution of foredeep-foreland basin systems (e.g., De Celles & Giles, 1996). In particular, extension in Natih A time can be interpreted to be triggered by forebulge formation and related uplift and erosion, whereas faulting in Muti and lower Fiqa times can be related to the subsequent foredeep formation and rapid subsidence (Robertson, 1987; Warburton et al., 1990; Breton et al., 2004). Ongoing plate convergence eventually led to the formation of the Oman Mountains foreland thrust-fold belt, mainly in Oligocene to Pliocene times, which includes further shortening in the Salakh Arc (Searle, 1985; Mount et al., 1998; Poupeau et al., 1998; Al-Lazki et al., 2002; Filbrandt et al., 2006; Fournier et al., 2006; Agard et al., 2010).

The presence of widespread NE-SW left-lateral strike-slip faulting associated with E-W fold growth is not a typical feature of positive fault inversion and cylindrical folding in general. Attempts to explain NE-SW strike-slip and N-S extensional faulting by progressive arching in a foreland salient fail to successfully reproduce actual shear senses. In fact, southward amplification of the Salakh Arc would have been mainly accommodated by right-lateral rather than left-lateral strike-slip faulting in Qusaybah (e.g., Hanna & Smewing, 1996).

Involvement of the Ara evaporites and the presence of a variable thickness overburden may provide the appropriate framework for explaining the peculiar fault-fold kinematics described in the Jabal Qusaybah anticline. Sandbox analogue modeling of thrust wedge de-

velopment above a salt layer characterized by very gentle, foredeep-parallel tapering of the overburden showed that the combination of tectonic and gravitational forces triggered development of a complex fault pattern where partitioned transpression and transtension favoured oblique extrusion of the frontal thrust sheet along fault zones striking at about 45° from the fold axial trend (Storti et al., 2007, Figure 4.19d, e). Application of this scenario to the toe region of the Oman Mountains thrust wedge implies: (i) a thicker sedimentary or tectonic overburden of the Ara evaporites in the eastern part of the Salakh Arc area; (ii) pinching of the Ara evaporites along the outer edge of the Salakh Arc; (iii) occurrence of NE-SW left-lateral strike-slip fault zones all across the Salakh Arc and not limited to the Jabal Qusaybah anticline. Condition (ii) is satisfied because the Salakh Arc region locates at the eastern termination of the Fahud salt basin (e.g., Filbrandt et al., 2006, Figure 4.1). Conditions (iii) is satisfied as well because NE-SW trending fractures, including left-lateral strike-slip fault zones, are described in other anticlines of the Salakh Arc (e.g., Bazalgette, 2004; De Keijzer et al., 2007). Condition (i) may have occurred also because crustal accretion from both the NNE (Samail Ophiolite in the Oman Mountains) and ESE (Masirah Ophiolite in East Oman) might have induced foredeep preferential subsidence towards the east. Moreover, onset of sinistral faulting was possibly favoured by inherited NE-SW trending weakness zones in the basement (e.g., Filbrandt et al., 2006).

Involvement of the Ara evaporites in the subsurface architecture of the Jabal Qusaybah anticline is not supported by the proposed interpretation of the seismic profile in Figure 4.17. This apparent contradiction could be due the lack of a diagnostic seismic facies for the evaporites, which are very old in this case, and/or by their masking by the strong noise in the anticlinal core caused by the horizontally variable velocity field across the fold. It is worth noting, however, that the seismic profile is located in the western pericline of the fold, where evaporites may have migrated towards the anticlinal core. The widespread occurrence of severely dilated fault zones and related in-fill by very large calcite crystals and aggregates has been described in Jabal Madar, where a salt diapir is well documented (e.g., Immenhauser et al., 2007; Claringbould et al., 2013), a related fold located ~ 110 km to the east of Jabal Qusaybah. The isotopic signatures of these crystals in Jabal Madar (Immenhauser et al., 2007) and in one of the very large crystal populations in Jabal Qusaybah (Figure 4.16b) are very similar. This evidence supports the inference of a role of the Ara evaporites in the growth of the Jabal Qusaybah anticline and helps explaining development of the along strike axial bulge in the central sector of the crestal region, possibly induced by late-stage salt migration (Figure 4.19c).

The proposed evolution of the Jabal Qusaybah Anticline contrasts with the tectonic architecture described in the foreland region to the

south of the Salakh Arc, which is dominated by NNW-SSE and WNW-ESE strike-slip fault arrays indicating a NW-SE trending, near horizontal maximum compressive stress during Late Santonian to Campanian times (Filbrandt et al., 2006). Similar fault trends are not widespread in Jabal Qusaybah, where strike-slip motions mostly occurred along NE-SW trending fault zones. The latter are almost perpendicular to the regional compression inferred to the south and this makes their re-activation quite unlikely. Similar considerations hold for positive inversion of the ~ E-W striking extensional master fault zone in the core of the Jabal Qusaybah anticline, which should have been affected by strong transpressional deformations. To reconcile the geodynamic settings to the north and to the south of the Salakh Arc, respectively, we may assume that the latter developed from an intraforeland crustal weakness zone suitable to provide an effective mechanical decoupling between the ~ N-S near horizontal maximum compressive stress responsible of the growth of the Oman Mountains since Early Turonian times, and the onset to the south of a NW-SE near horizontal maximum compressive stress associated with the emplacement of the Masirah Ophiolite in Late Maastrichtian - Paleocene times (e.g., Beurrier, 1987; Schreurs & Immenhauser, 1999).

4.7 CONCLUSIONS

We performed a structural and petrographical-geochemical study of deformation structures exposed in the Jabal Qusaybah anticline, at the western termination of the Salakh Arc, a major salient at the toe of the Oman Mountains thrust wedge. Field and laboratory data have been integrated with the analysis of a reflection seismic profile cutting through the western periclinal termination of the fold. The results of our work can be summarized in the following points.

1. Both field and reflection seismic data indicate that the anticline developed by positive inversion of a north-dipping extensional fault zone that formed in early Turonian times during deposition of Natih A sediments.
2. The deformation pattern characterizing the doubly-plunging, E-W trending Jabal Qusaybah anticline includes NE-SW left-lateral strike-slip and N-S extensional fault zones. Petrographical features, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ stable isotope signatures of differentiated calcite vein infills, and the geometric, kinematic, and overprinting relations of calcite veins indicate that folding and faulting occurred coevally during the growth of the anticline.
3. In the central sector of the anticlinal crest the fold axis has a gentle along strike bump. There, N-S striking extensional fault zones sub-perpendicular to the fold axis are abundant and abut the NE-SW left-lateral strike-slip fault zones. Both strike-slip

and extensional fault zones show widespread evidence for substantial dilation in the form of infilling by large calcite crystals and aggregates. This is particularly outstanding along the N-S extensional fault zones where the magnitude of fault core dilation frequently exceeds one meter.

4. A plausible kinematic pathway that could explain the coexistence of such unusual synfolding noncylindrical deformation patterns implies motion of the underlying Ara evaporites in contraction under a foredeep-parallel tapering overburden wedge of sedimentary and/or tectonic origin. This geometry would add a component of south-eastward extrusion to the evolution of the Salakh Arc, thus triggering NE-SW left-lateral strike-slip faulting, and would have favoured longitudinal axial bulging of the anticline accompanied by generalized dilation in the central region of the crest.
5. This work illustrates how complex deformation patterns can be in fault-related anticlines that develop near or within curved segments of major thrusts. Conventional templates for folding-related fracturing do not predict the occurrence of either the N-S extensional or NE-SW strike slip fault systems. Similar atypical fault and fold patterns could be expected in structures at the toe of thrust wedges, and so should be incorporated into conceptual models of those structures. We consequently stress the importance of studying appropriate field analogues of fault-related folds constrained using a structural and petrographical-geochemical multidisciplinary approach. Understanding the timing of the fault, fold and fracture system growth would not have been possible without this integrated approach.

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VAL DI LIMA FAULT SYSTEM

This chapter is presented in the form of a manuscript submitted to the *Journal of Structural Geology*. It describes the interactions between large-scale compressional and extensional structures developed in the Val di Lima region of the Northern Apennines thrust wedge. The area of interest has been cyclically affected by compressional and extensional deformations, and the resulting deformation structures mutually overprint to give rise to a complex 3D architecture.

SUBMITTED ARTICLE

Extensional deformation structures within a convergent orogen: The Val di Lima low-angle normal fault system (Northern Apennines, Italy).

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EXTENSIONAL DEFORMATION STRUCTURES WITHIN A CONVERGENT OROGEN: THE VAL DI LIMA LOW-ANGLE NORMAL FAULT SYSTEM (NORTHERN APENNINES, ITALY).

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ABSTRACT

A low-angle extensional fault system affecting the non metamorphic rocks of the carbonate dominated Tuscan succession is exposed in the Lima valley (Northern Apennines, Italy). This fault system affects the right-side-up limb of a kilometric-scale recumbent isoclinal anticline and is, in turn, affected by superimposed folding and late-tectonic high-angle extensional faulting.

The architecture of the low-angle fault system has been investigated through detailed structural mapping and damage zone characterization. Pressure-depth conditions and paleofluid evolution of the fault system have been studied through microstructural, mineralogical, petrographic, fluid inclusion and stable isotope analyses. Our results show that the low-angle fault system was active during exhumation of the Tuscan succession at about 180°C and 5 km depth, with the involvement of low-salinity fluids. Within this temperature - depth framework, the fault zone architecture shows important differences related to the different lithologies involved in the fault system and to the role played by the fluids during deformation. In places, footwall overpressuring influenced active deformation mechanisms and favored shear localization.

Our observations indicate that extensional structures affected the central sector of the Northern Apennines thrust wedge during the orogenic contractional history, modifying the fluid circulation through the upper crust and influencing its mechanical behavior.

5.1 INTRODUCTION

Thrust wedges are characterized by complex evolutionary histories that originated from the interactions between tectonics, gravitational

forces, and surface processes (e.g., Elliott, 1976; Chapple, 1978; Storti & McClay, 1995; Mugnier et al., 1997; Konstantinovskaya & Malavieille, 2005; Bonnet et al., 2007; Malavieille, 2010). They are typically characterized by the development of contractional structures associated with processes such as underplating, thrusting, and folding. Nevertheless, important stages of wedge extension play a major role in the long-term evolution of thrust wedges, maintaining their cross-sectional slope in the critical taper range (e.g., Davis et al., 1983; Platt, 1986). Indeed, extensional deformations may affect sectors of a thrust wedge while other sectors are dominated by contraction, in a non-systematic way. Accordingly, thrust wedge sectors can be cyclically affected by contractional and extensional deformations overprinting each other and producing complex structural architectures.

When extensional deformations occur in the upper portion of thrust wedges, complex systems of brittle extensional fault zones develop which influence fluid circulation through the upper crust by producing a complex network of hydraulic conduits, barriers, or mixed conduit-barrier systems (Caine et al., 1996). The latter are strongly controlled by fault system geometry, fault zone architecture and rock-type. In particular, extensional fault zones are proven to provide effective pathways allowing deep infiltration of surface-derived marine or meteoric fluids, down to their basal low-angle detachments (Wickham & Taylor, 1987; Fricke et al., 1992; Morrison, 1994; Morrison & Anderson, 1998; Bebout et al., 2001; Verhaert et al., 2004, 2009; Mulch et al., 2004, 2006; Person et al., 2007; Ingebritsen & Manning, 2010; Gébelin et al., 2011; Gottardi et al., 2011). These fluids, in turn, exert a strong influence on the mechanical behavior of thrust wedges either by reducing the effective normal stress at depth or triggering the formation of hydrous clay minerals lowering the frictional properties of fault zones (Hubbert & Rubey, 1959; Davis et al., 1983; Dahlen et al., 1984; Sibson et al., 1988; Dahlen, 1990; Moore & Vrolijk, 1992; Kirschner & Kennedy, 2001; Collettini & Holdsworth, 2004; Fagereng et al., 2010; Lacroix et al., 2013).

In this paper, we describe an example of mutually overprinting contractional and extensional structures from the Northern Apennines (Italy), characterized by: i) a kilometric-scale recumbent isoclinal fold; ii) the Val di Lima low-angle extensional fault system affecting the right-side-up limb of this fold; iii) a system of open folds and a foreland-vergent anticline deforming the fault system; iv) high-angle extensional and strike-slip fault zones overprinting all previous structures. In particular, we focus on the low-angle fault system and use a multidisciplinary approach to investigate the paleofluid evolution of this structure.

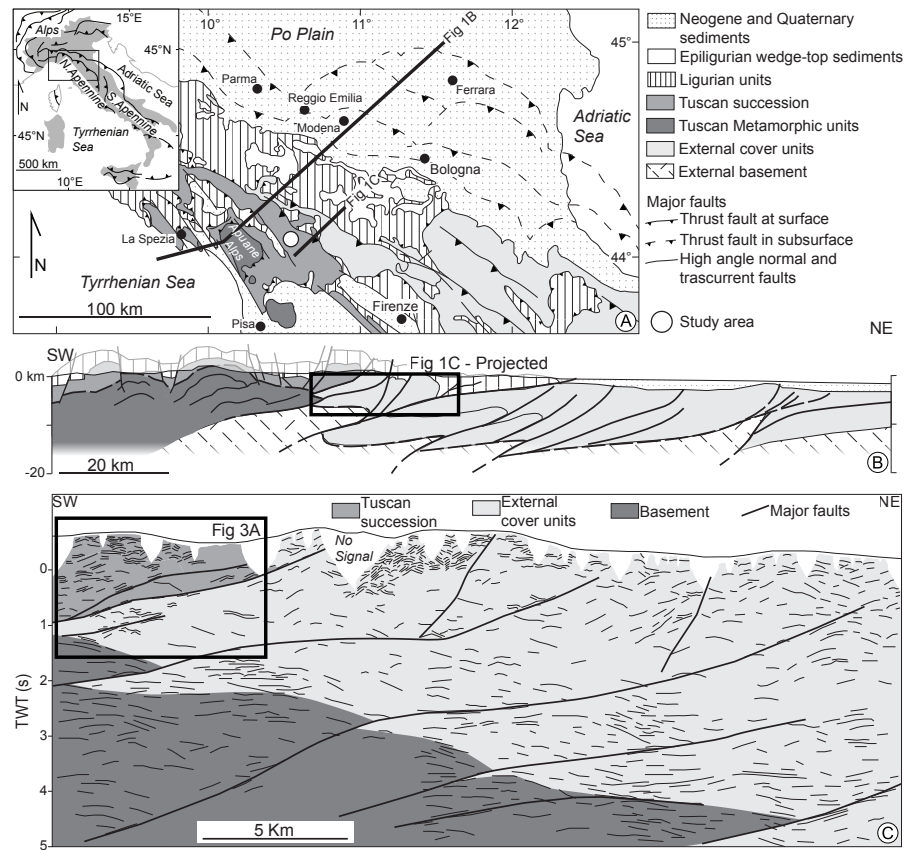


Figure 5.1: Regional-scale geological setting. a) Tectonic map of the Northern Apennines (Molli, 2008); the white dot indicates the location of the study area. b) Schematic crustal-scale cross-section of the Apenninic thrust wedge, located approximately 25 km to the NW of the study area (Molli, 2008). c) Line drawing and geological interpretation of a commercial seismic profile recorded close to the study area.

5.2 GEOLOGICAL OUTLINE

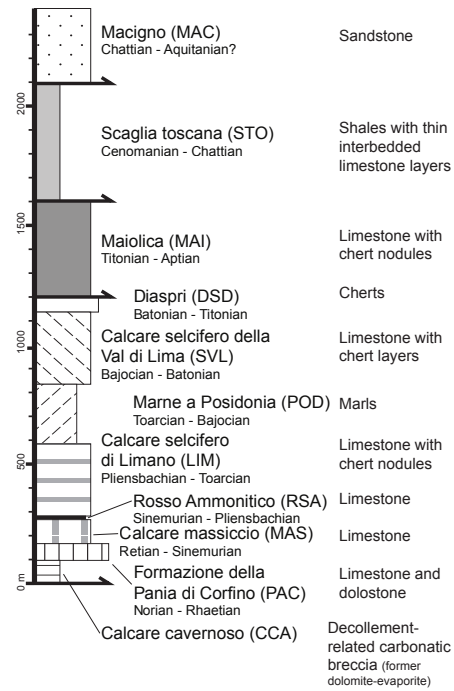
The Apenninic segment of the Alpine Orogenic system (Figure 5.1) has been growing on top of the retreating subduction of the Adria plate below the European plate since Oligocene times (Malinverno & Ryan, 1986; Royden, 1988; Dewey et al., 1989; Patacca et al., 1990; Doglioni, 1991). The remnants of the Ligurian ocean and of the thinned passive margin of the Adria Plate are accreted in the Northern Apennines thrust wedge (Boccaletti et al., 1971; Elter & Pertusati, 1973). From bottom to top the nappe stack includes: i) the mainly carbonate Tuscan succession and its crystalline basement, deformed under very low- to low-grade metamorphic conditions, ii) the non-metamorphic to very low-grade metamorphic ophiolite-bearing Ligurian succession, and iii) the Epiligurian wedge-top sediments (Elter, 1975). The first-order nappe pile underwent significant tectonic thinning by low-angle extensional faulting (Carmignani & Kligfield, 1990), particularly in the southern portion of the inner Northern Apennines, where they are overprinted by out-of-sequence thrusting (Storti, 1995; Ar-

gnani, 2002; Molli, 2008; Boccaletti et al., 2011) and by further extensional deformations that triggered also upper Miocene to Pleistocene magmatic activity (e.g., Elter, 1975; Carmignani et al., 1995; Storti, 1995; Barchi et al., 1998; Decandia et al., 1998; Jolivet et al., 1998; Brogi, 2004, 2008; Collettini et al., 2006b). The deformation conditions of the Tuscan succession within the Apenninic thrust wedge have been constrained through different methodologies including vitrinite reflectance (Reutter et al., 1983), stable isotopes (Carter & Dworkin, 1990), fission track on apatite (Abbate et al., 1994; Zattin et al., 2002; Balestrieri et al., 2003) and zircon (Bernet et al., 2004; Fellin et al., 2007), fluid inclusion microthermometry (Montomoli et al., 2001; Montomoli, 2002), illite crystallinity (Cerrina Feroni et al., 1983; Carosi et al., 2003), calcite-dolomite geothermometry (Carosi et al., 2003) and TEM microstructure analysis (Molli et al., 2011). The maximum tectonic burial of the Tuscan succession, related to the first thrusting event, took place at about 15 - 10 Ma, with the deepest conditions at ~ 7 km depth and 200-250°C, recorded in the internal portion of the Northern Apennines (La Spezia area; Molli et al., 2011), and progressively shallower conditions towards the east (Fellin et al., 2007). The nominal geothermal gradients corresponding to the different paleotemperatures calculated by the different authors range between 18 and 41°C/km (Molli et al., 2011, and references therein) with an average value of $31 \pm 4^\circ\text{C}/\text{km}$.

In the Lima valley, all formations belonging to the Tuscan succession are exposed (Figure 5.2; Botti et al., 2010, and references therein), from the Calcare Cavernoso basal tectonic breccia, to the ~ 2 km thick Triassic to Eocene passive margin sedimentary sequence, up to the Chattian – Aquitanian? Macigno foredeep siliciclastic deposits (Costa et al., 1992). From a structural point of view, the Tuscan succession is locally involved in a kilometric-scale isoclinal recumbent fold facing toward the northeast, with the reverse limb outcropping in the bottom of the Lima valley and the sub-horizontal axial surface localized in the Calcare Cavernoso Formation (Figure 5.3; Zaccagna, 1935; Baldacci et al., 1967). The Val di Lima low angle fault system affected the right-side-up limb of the isoclinal fold. Both the recumbent fold and the low-angle fault system are refolded in a NE-verging asymmetric anticline with a vertical forelimb (Figure 5.3C). Finally, high-angle extensional and strike-slip fault zones overprint the previously formed structures.

The Val di Lima low angle fault system has been investigated by many authors during the past fifty years, proposing different interpretations: i) extensional structure developed in the domain interposed between two vicariant NW-SE strike-slip high-angle faults (Bellincioni, 1959); ii) a thrust sheet affected by multiple folding (Pertusati et al., 1977; Baldacci et al., 1982); iii) a complex reverse fault with flat segments localized along few low-angle detachment levels and local down-section trajectories (Fazzuoli et al., 1998); iv) a SW vergent

Figure 5.2: Simplified stratigraphic column of the Tuscan succession exposed in the Lima Valley, with indicative thickness of the various formations (after Botti et al., 2010).



extensional fault system partially reactivated during successive NE-vergent thrusting (Calò, 1998); v) a splay fault zone of a regional-scale low-angle extensional fault system (Botti et al., 2010).

5.3 METHODS AND ANALYTICAL TECHNIQUES

To reconstruct the original kinematics of the Val di Lima low-angle fault system, superimposed folding and high-angle faulting have been restored on geologic cross-sections and the tectonic transport directions have been recomputed after rotating the mean attitude of the slip surfaces to the horizontal in all structural sites. Stereographic projections and structural data rotations were made by the Daisy 3 software (Salvini et al., 1999).

With regard to laboratory analyses, each rock sample was cut to obtain two “mirrored” slabs, one of which was polished and stained with alizarin red and potassium ferricyanide to discriminate different carbonate minerals such as calcite, dolomite and their iron-rich phases. From the other slab, thin-section and wafers were prepared and thin sections were examined by conventional and cold cathodoluminescence (CL) petrography. The latter was carried out with a Technosyn 8200 MK II stage operated at 8kV, 200 μ A gun current.

X-ray diffraction mineralogical analyses (XRD) were performed with a Philips PW 1710 automatic diffractometer, operated at 40 kV and 20 mA. Preliminary XRD analyses in the interval $4^\circ < 2\theta < 60^\circ$ were performed to identify the mineralogical phases. XRF analysis of major elements was carried out on fused glass discs prepared by mixing powdered rock samples and lithium tetraborate according to the methodology proposed by Tamponi et al. (2002). Chemical

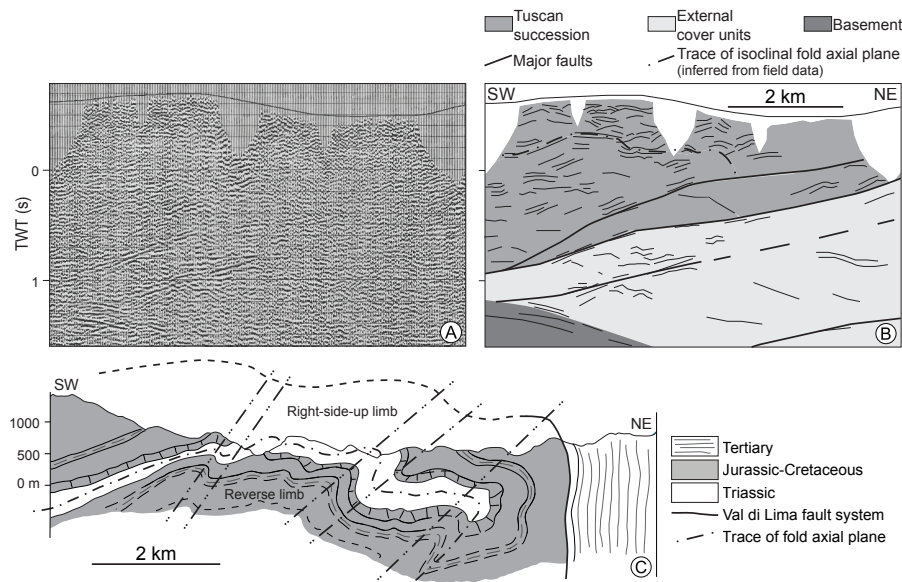


Figure 5.3: Geological framework of the Lima valley region based on surface and subsurface data. A) Close up view of the commercial seismic profile in Figure 5.1C. B) Line drawing and interpretation of Figure 5.3A. C) Schematic cross-section showing the main structural features of the Lima valley recumbent fold: the anticline has horizontal limbs and is affected by the Val di Lima detachment, located at the top of the Triassic-Cretaceous carbonates. This structural architecture is overprinted by younger open folds (modified from Baldacci et al., 1992).

concentrations have been calculated from fluorescence lines of major elements and expressed in relative percentages of the respective oxides. Quantitative estimation of the mineralogical phases constituting each samples were then computed from the chemical concentrations according to the methodology proposed by Leoni et al. (1989). To investigate clay mineralogy in more detail, the $< 2 \mu\text{m}$ grain-size fraction was separated by sedimentation from sample powders, and specific XRD analysis were performed on oriented slides in the interval $4^\circ < 2\theta < 15^\circ$. For each sample, multiple $< 2 \mu\text{m}$ slides were prepared and subjected to different treatments, namely: i) K^+ saturated and air dried (K^+); ii) Mg^{2+} saturated and air dried (Mg^{2+}); iii) Mg^{2+} saturated and ethylene glycol solvated (Mg^{2+} EG).

Fluid inclusions were analyzed in doubly polished thick sections (wafers) which were mapped with a standard petrographic microscope. The wafers were subsequently broken into smaller pieces (chips). Whenever possible, the fluid inclusion assemblage (FIA) approach was followed (Goldstein & Reynolds, 1994). These fluid inclusion assemblages do not show any relationship to fractures, occur randomly in the sample and are considered to represent the trapping or recrystallization conditions of the minerals analyzed. Each FIA is composed of few inclusions (generally 5 to 10) considered to be representative of the same trapping event. Few isolated inclusions were measured as well, but only if many of them were found in the same vein. Sam-

ples with only few isolated inclusions (less than 5 per vein) were not further considered. A Linkam THMSG600 heating-cooling stage was used for microthermometric analyses and weekly calibrated with SynFlinC synthetic standards. In order to avoid any artificial stretching of the inclusions, heating was always performed before freezing and each chip was heated only once. Analyzed inclusions are 2 to 5 μm in size and two-phase at room temperature (aqueous liquid + vapor bubble). Vapor bubble size range from 10 to 20% in volume in analyzed inclusions; inclusions with bigger bubbles are rare and were not further considered. Three temperatures were routinely measured: homogenization to the liquid phase temperature (T_h), indicative of the minimum trapping temperature of the inclusion; first ice melting temperature (T_{fm}), indicative of the fluid system; final ice melting temperature ($T_{m\ ice}$), inversely proportional to the salt content of the aqueous phase. Fluid salinity was computed from measured $T_{m\ ice}$ values using the software FLUIDS (Bakker, 2003). Pressure correction was applied to the T_h values to determine the actual trapping conditions (temperature and pressure) of each FIA. The trapping conditions were defined by the intersection of the fluid pressure gradient (computed from the estimated geothermal gradient) and the appropriate isochore (computed from the salinity and T_h of each group of inclusions). Some inclusions, biphasic at room temperature, showed a metastable behavior during re-heating after being frozen: their bubble returned to the original pre-freezing size at temperatures above 0°C despite the slow heating rate (0.1°C/minute). Such inclusions are plotted all together with $T_{m\ ice} = m$, without reporting the actual measured $T_{m\ ice}$.

Stable isotope analysis was done on microsamples of veins and host rocks. Microsamples were drilled from the stained slabs (after removal of the material altered by staining) with a dentist drill. Carbonate powders reacted with 100% phosphoric acid at 70°C using a Gasbench II connected to a Thermo Finnigan Five Plus mass spectrometer. All values are reported in per mil (‰) relative to the Vienna PDB standard. Reproducibility was checked by replicate analysis of selected samples and laboratory standards and is better than ± 0.07 and ± 0.09 ‰ for oxygen and carbon isotope respectively. Isotope fractionation curves were computed with the formula

$$10^3 \ln \alpha = \frac{A 10^6}{T^2} + B \quad (5.1)$$

using $A = 2.78$, $B = -2.89$ (Friedman & O'Neil, 1977) and $A = 3.14$, $B = -2$ (Land, 1983) for calcite and dolomite, respectively.

5.4 STRUCTURAL ARCHITECTURE OF THE VAL DI LIMA DETACHMENT FAULT

The Val di Lima fault system has a complex three-dimensional architecture (Figures 5.4 and 5.5), characterized by staircase geometry with

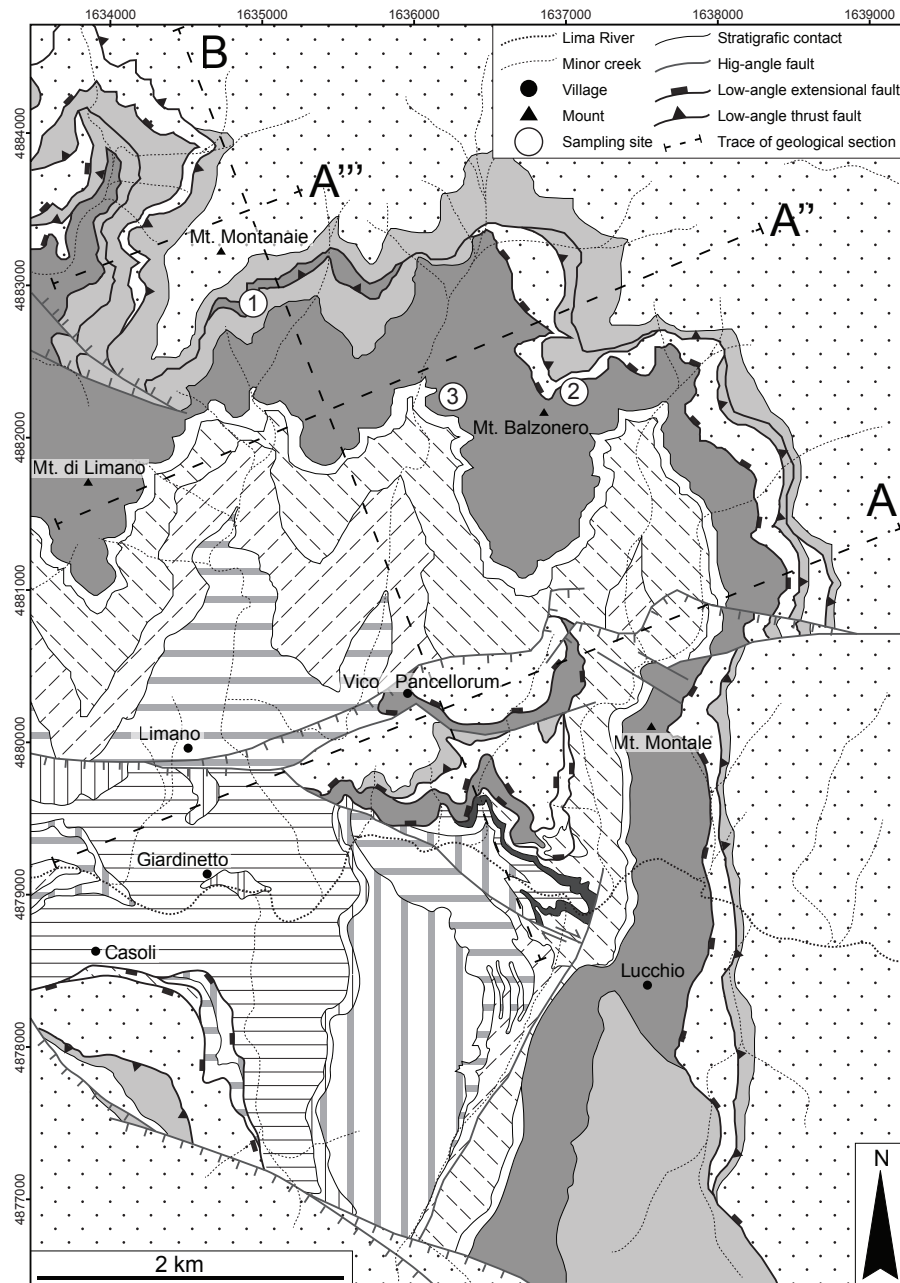


Figure 5.4: Simplified geological map of the study area (modified from Servizio Geologico d'Italia, 2012).

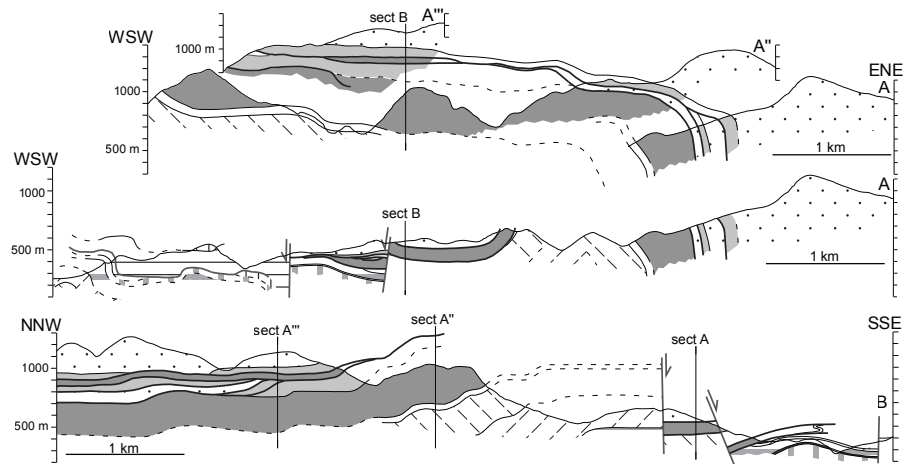


Figure 5.5: Geologic cross-sections across the study area, highlighting the present-day attitude of the Val di Lima low-angle fault system (see Figure 5.4 for section locations).

major flat segments connected by minor splays with anastomosed lateral terminations. Within this framework, remarkable local variations exist in the different sectors of the study area. In the northwestern sector (west of Mt. Montanaie), a thin slice of Macigno is tectonically embedded in between the Scaglia shale. On the eastern prosecution of such deformation zone (south of Mt. Montanaie), two discrete flat-lying slip surfaces originate from a branch point and delimit a thin (25 - 50 m) lens of Maiolica limestone, which tectonically overlies the Scaglia Formation (Figure 5.4). In the eastern sector (Mt. Balzonero – Mt. Montale - Lucchio village), the fault system is composed of two parallel major segments bounding a ~ 100 m thick, 6 km long slice of Macigno sandstone (Figures 5.4 and 5.5), which directly lies on the Maiolica Formation via the complete excision of the Scaglia Formation. The latter, in turn, tectonically overlies the Macigno slice and is internally thinned by subsidiary splay fault zones. In the central sector of the study area (in the proximity of Vico Pancellorum village), the fault system is composed of three major low-angle segments. From bottom to top, they tectonically juxtapose i) a ~ 100 m thick slice of folded Maiolica, Diaspri and Calcare Selcifero della Val di Lima Formations, ii) a ~ 50 m thick slice of Macigno sandstone, iii) an upper slice of Scaglia shale (Figures 5.4 and 5.5). In the southeastern sector of the study area (in the proximity of Casoli village), the fault system is composed of two main segments. The upper one causes the juxtaposition of the upper portion of Scaglia shale above a ~ 200 m thick Macigno sandstone slice. The lower segment juxtaposes the Macigno slice onto the Calcare Cavernoso Formation via the excision of all the intervening formations, apart from few small slices delimited by subsidiary splay fault zones.

Four hundred and forty structural data were collected along the exposed segments of the low-angle fault system in order to constrain

their kinematics (Figure 5.6). Regardless of the local attitude of deformation structures, which shows broad variations in the different sectors due to involvement in successive deformation events, the transport direction is almost constantly toward NE, with slight variations in the N020E - N060E range. The recomputed original tectonic transport directions after restoration of superimposed folding (big arrows on the map of Figure 5.6) slightly differ from the present-day ones.

5.5 FAULT ROCKS AND DAMAGE ZONE ARCHITECTURE

Fault rocks were studied in two localities where outcrop conditions allowed a good observation of both the main slip surface and the damage zone architecture (Site 1 and Site 2 in Figure 5.4). For comparison purpose, few samples of protolith rocks were also collected (Site 3 in Figure 5.4). Noteworthy, in the selected outcrops three different lithologies are involved in the Val di Lima low-angle fault system, both in the hanging wall and in the footwall, providing the opportunity to study the influence of rock type on fluid flow and deformation structure development in low-angle fault damage zones.

5.5.1 *Structural analysis*

5.5.1.1 *Site 1*

Site 1 is located south-east of Mt. Montanaie, where a slice of Maiolica limestone is thrust over the Scaglia shales. The Maiolica slice is delimited by bedding-parallel shear zones, with the bottom one being particularly well exposed. It is gently dipping toward the NW and shows marked slickenlines and centimeter-scale grooves oriented toward the NE. Few high-angle subsidiary faults are present in the hangingwall damage zone. They are NW-SE oriented and the sole down on the main slip surface. In the hanging wall of such a basal fault zone, the Maiolica limestone shows a meter-scale damage zone characterized by a breccia texture with the presence of abundant chert nodules (Figures 5.7A and 5.8A). In the limestone, two vein systems are present: one is oriented NE-SW (parallel to the tectonic transport direction) and the other one is oriented NW-SE (orthogonal to the tectonic transport direction). In chert nodules, veins are much more abundant than in the limestone, and they all belong to the NW-SE system. Tensile veins are filled by calcite, and sample staining indicates high iron content for both the veins and the limestone host. Vein aperture is generally few millimeters and texture is equigranular. Calcite crystals in the veins show abundant type I internal deformation twinning indicative of low-temperature deformation (Burkhard, 1993). Some crystals show only one twinning system, while other crystals show two bisecting twinning systems; spacing between the adjacent twinning planes is generally $\sim 10\mu\text{m}$. In cathodolumines-

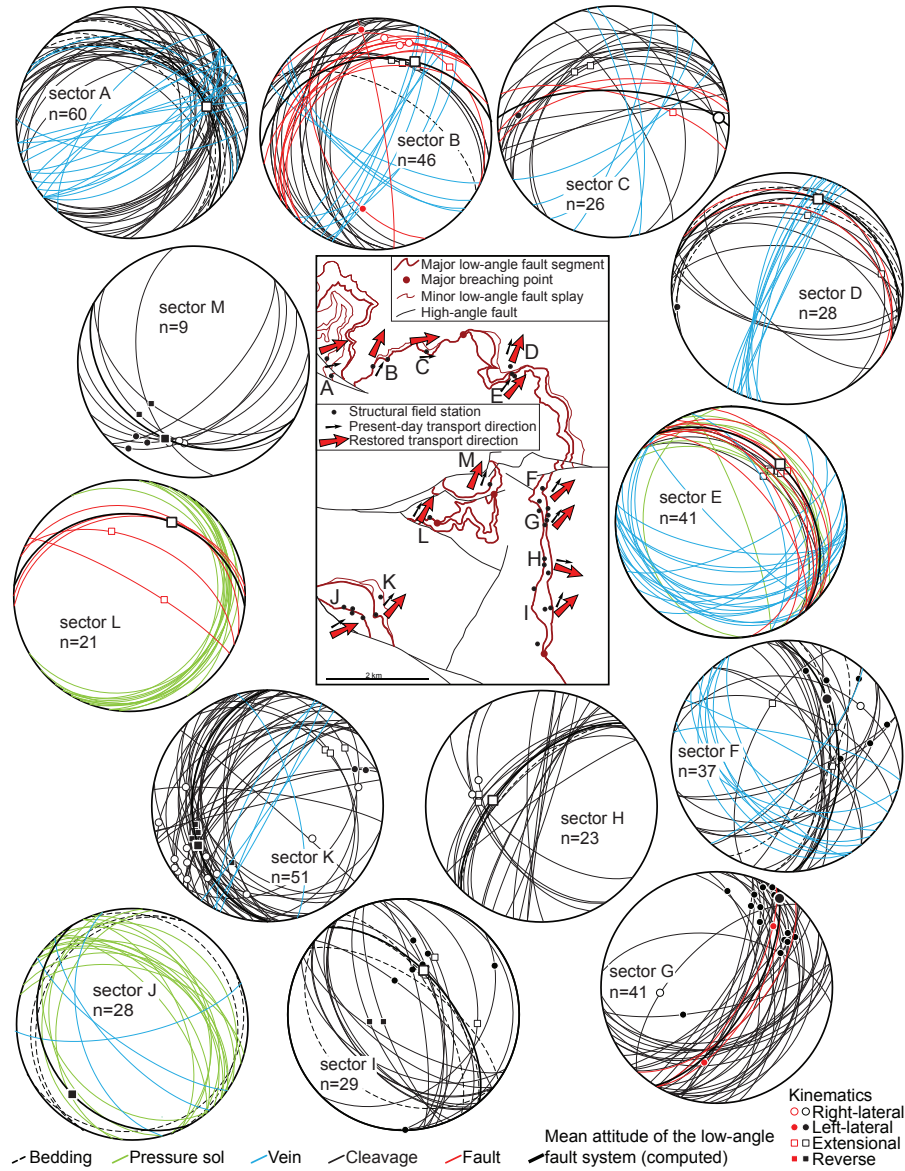


Figure 5.6: Stereographic projections (Wulff diagram, lower hemisphere) of the structural data collected in the study area and computed mean attitude/transport direction of the low angle fault system in each sector. The azimuth of the computed transport direction is also shown in map view, together with the restored transport direction (see text for details).

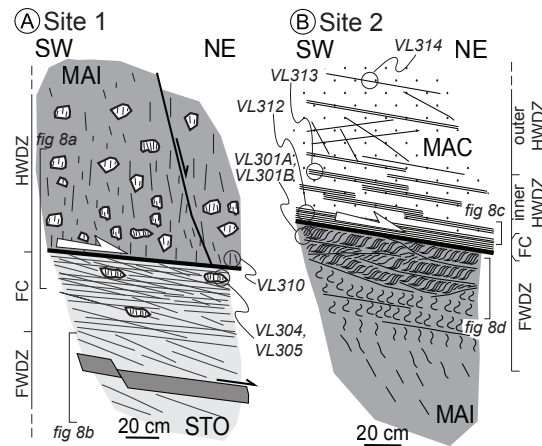


Figure 5.7: Schematic cross sections of the low angle fault system damage-zone architecture in (A) Site 1 and (B) Site 2 (see Figure 5.4 for location). Dotted: Macigno sandstone (MAC); light gray: Scaglia shales (STO); dark gray: Maiolica limestone (MAI) and competent layers in Scaglia; white: chert nodules; FWDZ: footwall damage zone; FC: fault core; HWDZ: hangingwall damage zone. The location of outcrop photos and the position of key samples for microstructural, mineralogical and fluid inclusion analysis are indicated in *italics*.

cence microscopy, the host rock is bright while calcite veins are dull with evidences of bright reopening/recrystallization (Figure 5.9A).

The fault core of the basal tectonic contact is marked by a few tens of centimeters thick gouge layer. The gouge is mostly developed at the expense of the underlying Scaglia Formation but some chert nodule remnants, most likely derived from the overlying Maiolica, are also present. Fault core chert nodules are crosscut by NW-SE tensile veins, similarly to chert nodules in the hangingwall damage zone, but in this case they show a mixed quartz / calcite infilling. Quartz crystals are clear, equigranular, do not show internal deformation and are 100 to 300 μm in size. Calcite, on the contrary, is completely recrystallized and can be either brown colored or light colored in plane light; in the former case the calcite is very bright luminescent, while the latter calcite is dull luminescent (Figure 5.9B).

In the Scaglia formation at the footwall of the basal fault zone, deformation is delocalized in a several tens of meters thick damage zone. A very intense fault parallel foliation developed in the shale layers, with a low-angle attitude with respect to bedding. The more competent marly limestone layers form shear lenses with minor internal deformation, typically bounded by polished shear surfaces with evident NE-oriented slickenlines (Figure 5.8B, inset). In places, the competent layers are cut by centimeter-scale subsidiary high-angle NE-verging extensional faults that are consistent with the general top-to-NE kinematics of the whole low-angle fault system (Figures 5.7A and 5.8B).

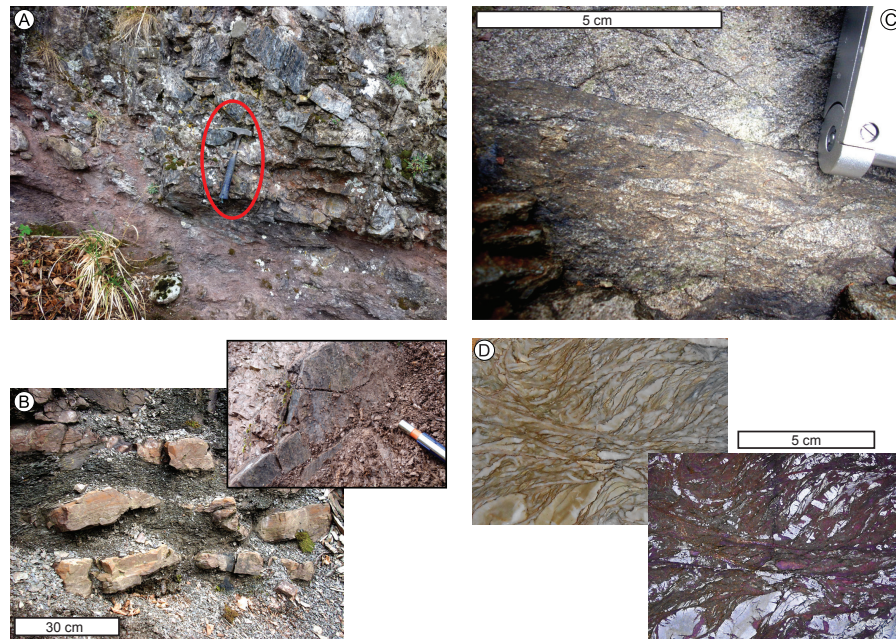


Figure 5.8: Fault rock pictures. A) Site 1: brecciated Maiolica limestone in site 1 hangingwall damage zone (hammer for scale). B) Site 1: delocalized deformation in the Scaglia Formation few meters away from the main slip surface. The shale layers show a pervasive low angle foliation and the more competent limestone layers are cross-cut by small scale high-angle subsidiary faults. The inset shows the polished surface of a competent layer, with NE-oriented slickenlines (pencil for scale). C) Site 2: detail of the fault core and transition to the hangingwall damage zone in the Macigno sandstone. D) Site 2 footwall damage zone: polished (left) and stained (right) rock slab of Maiolica carbonates showing abundant calcite veins, deformed in a S-C pattern (sample VL301A). Staining indicates iron-poor dolomitic host rock (light gray) and iron-rich calcite veins (purple).

5.5.1.2 Site 2

At site 2, the Macigno sandstone is in direct contact with the Maiolica limestone via the complete excision of the Scaglia Formation. A very thin layer (up to 5 cm thick) of foliated cataclasite constitutes the fault core (Figure 5.7B). The foliated cataclasite has a matrix-supported texture with size reduction of both quartz and feldspar grains (Figure 5.10A). Abundant calcite microveins, with antitaxial infilling showing type I internal twinning (Burkhard, 1993), affects survivor quartz clasts and rock fragments. The matrix is fine grained and enriched in oxide and phyllosilicates. Different microstructures are observed, including asymmetric microlithons, synthetic and antithetic low angle shears, fault-parallel shears and deflected shears with S asymmetry. They are all consistent with a top-to-NE tectonic transport direction.

In the hangingwall of the main slip surface, the Macigno sandstone developed a meter-scale damage zone, which can be divided in 2 different sectors (Figure 5.7B). The inner damage zone, close to

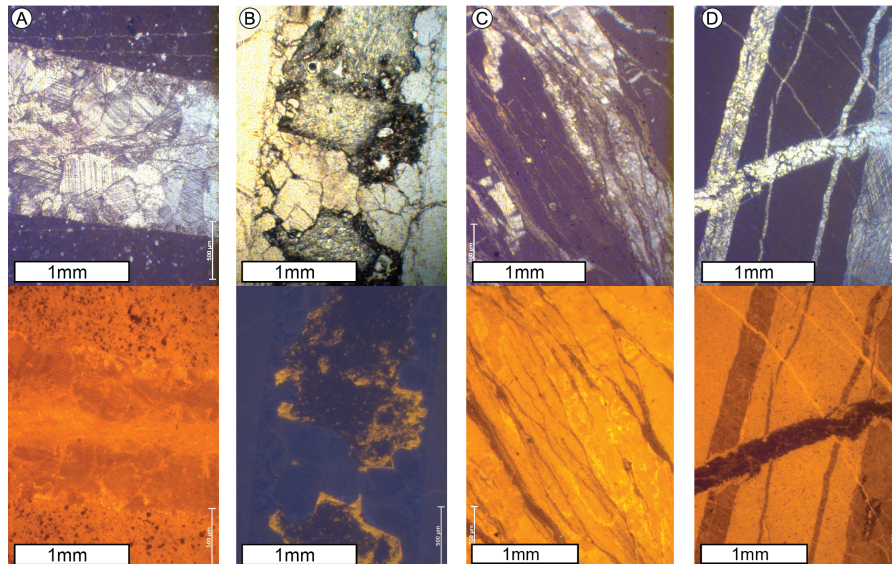


Figure 5.9: Microphotographs (MP) of fault rocks under plane light (upper MP) and cathodoluminescence view (lower MP), see Figure 5.7 for sample locations. A) Sample VL310: calcite vein in Maiolica limestone from Site 1 hangingwall damage zone. B) Sample VL305: mixed quartz – calcite vein in chert nodule in Site 1 fault core. C) Sample VL301A: S-shaped veins and host rock remnants in Maiolica from Site 2 footwall damage zone. D) Sample VL302: calcite veins in Maiolica limestone from the relatively undeformed Site 3, far from any major fault.

the main slip surface (~ 0 to 20 cm), includes lithons of poorly deformed sandstone separated by cm-thick domains of foliated rock (Figure 5.8C). Such deformed domains are, from a microstructural point of view, characterized by: i) thin, regularly spaced shear zones enriched in oxides and phyllosilicates, with fault-parallel to low-angle orientation; ii) millimetric microlithons of crushed breccias; iii) millimetric microlithons of relatively undeformed rock (Figure 5.10B). In the outer hangingwall damage zone, at greater distances from the main slip surface (~ 25 to 300 cm), large slices of Macigno sandstone are bounded by millimeter scale shear zones (Figure 5.7B). The undeformed Macigno sandstone has an isotropic clast-supported texture, characterized by angular clasts with bimodal grain-size distribution of 1.5 mm and 0.2 mm as the most representative grain sizes. The millimeter-thick shear zones are rectilinear, with sharp boundaries and fault parallel to low-angle orientation in the range of 10 – 15° . From a microstructural point of view they are characterized by a matrix-supported fabric with quartz and feldspar clasts smaller than those in the adjacent wall rock. Low-angle synthetic shears (R-orientation) and a deflected foliation with S asymmetry (P-foliation) also occur and indicate a top-to-NE tectonic transport direction (Figure 5.10C). Outside of the damage zone, only few minor structures are observed, such as: small high-angle subsidiary faults, tear faults

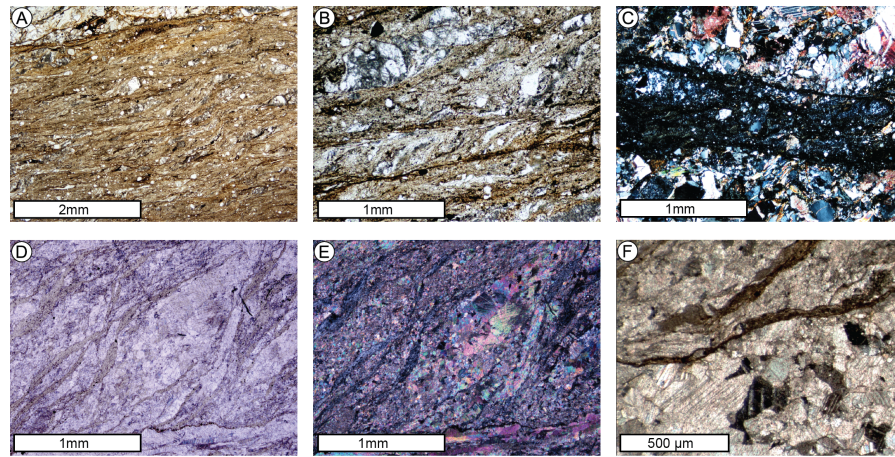


Figure 5.10: Microstructural aspects of fault rocks under optical microscope, see Figure 5.7 for sample locations (PV: Plane light view; CP: crossed polars). A) Sample VL312, foliated cataclasite in Site 2 fault core (PV). B) Sample VL313, Macigno sandstone in hangingwall damage zone: regularly spaced thin shear zones separating microlithons of crushed breccias and microlithons of relatively undeformed rock (PV). C) Sample VL314, Macigno sandstone far from the main slip surface: millimetric-thick shear zone cutting through undeformed sandstone (CP). D) Sample VL301A, Maiolica in footwall damage zone, close to the main slip surface: rock is almost made up of calcite veins. Two sets of veins are present, deformed in an S-C type shear fabric and showing type-I twinning (PV). E) Same as d (CP). F) Same as d, close up of the two vein systems (PV).

and fault-parallel shear planes presenting polished surfaces with slickenlines and anastomosed lateral terminations.

The Maiolica limestone in the footwall is characterized by abundant calcite veins, oriented at high angle with respect to the fault. Approaching to the main slip surface, vein spacing decreases and, in the same trend, vein shape changes from linear to curvilinear, arranged in tension gash arrays (Figure 5.7B). Close to the main slip surface (< 30 cm), the damage zone is characterized by peculiar features: i) veins are so closely spaced that they eventually constitute the majority of the rock volume, with host remnants being preserved only as centimetric-scale lithons; ii) vein curvature becomes very pronounced and with opposite asymmetry with respect to the tension gashes; iii) a second system of fault-parallel veins develops, characterized by linear shape and wider spacing. The combination of the two vein systems gives a well developed S-C shear fabric, consistent with a top-to-NE tectonic transport direction (Figure 5.7B). When this deformation structure developed the host rock lithons were made of iron-poor dolomite while veins were made of iron-rich calcite, as clearly shown by staining of the rock slabs (Figure 5.8D). Vein attitude, height and spacing were measured along various fault-parallel transects, at increasing distances from the master slip surface and analysis of mean values shows that: i) vein azimuth is quite constant with W-E orientation throughout the whole damage zone; ii) vein dip

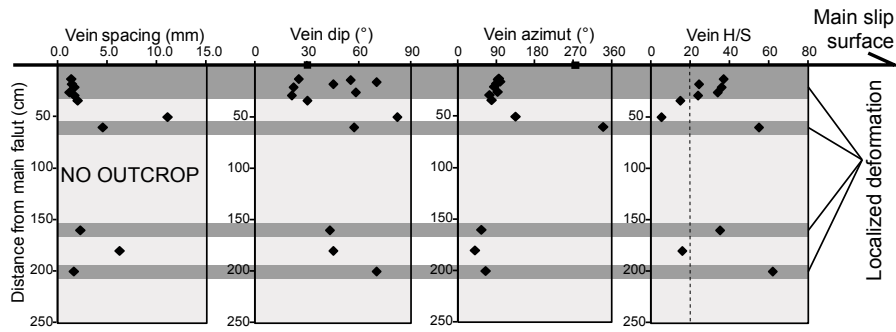


Figure 5.11: Quantitative characterization of the vein frequency in the damage zone affecting the Maiolica limestone at Site 2. Vein attributes have been measured along fault-parallel transects, 10 to 30 cm in length; for each transect the mean value of each attribute has been computed and plotted versus the distance between the transect and the main slip surface. Spacing values are measured as the spacing from the center of one vein to the center of next vein. Vein attitude is provided by the right-hand rule (Dip and Azimuth values). The H/S index is calculated as vein height divided by vein spacing.

is highly variable in the proximity of the main slip surface and is more regular at greater distances; iii) vein spacing is almost constant in the proximity of the master slip surface (from 0 to ~0.5 m distance) and is highly variable at greater distances, depending on deformation intensity (Figure 5.11). Vein spacing normalized versus vein height (H/S; Tavani et al., 2008) shows a strongly increasing trend, from about 4 at 0.5 m from the master slip surface, to about 38 very close to it. Comparable or even higher values occur in localized subsidiary shear zones (Figure 5.11). From a microstructural point of view, S-shaped high-angle veins are closely spaced (mean spacing ~0.2 mm) and thin (mean aperture ~0.05 mm) with very fine grained calcite infilling (Figure 5.10D and E). Some thicker veins with coarser grained calcite are also present. Fault parallel veins, on the other hand, have millimetric thickness, centimetric spacing, sharp boundaries with irregular fuzzy geometry and antitaxial calcite infilling (mean crystal size ~0.05 mm; Figure 5.10D and E). Veins of both systems are characterized by well developed type I twins (Figure 5.10F; Burkhard, 1993). Both dolomitic host rock and calcite veins are bright luminescent (Figure 5.9C).

5.5.1.3 Site 3

In the relatively undeformed Maiolica limestone, far from major fault zones, randomly oriented veins are present and slab staining indicates that both host rock and veins are made of iron-rich calcite. Combined structural and petrographic analysis allowed discriminating between different vein generations, with very variable apertures ranging from few millimeters, to few hundreds of μm , down to even smaller microveins. Veins usually shows equigranular crystals, but few bigger veins with elongated calcite are also present. Most of the

veins show abundant type I internal twinning (Burkhard, 1993). Some crystals show only one twinning system, while other crystals show two bisecting twinning systems; spacing between the adjacent twinning planes is generally $\sim 10 \mu\text{m}$. Twinned veins are cross-cut by rare younger veins showing no internal deformation. In cathodoluminescence microscopy, the host rock is bright luminescent, the twinned veins show various shades of dull luminescence, and the non twinned veins are non-luminescent (Figure 5.9D).

5.5.2 Mineralogical analysis

Chemical and mineralogical analyses were performed on two samples of Macigno sandstone from the Site 2 fault zone. Samples are representative of: i) the foliated cataclasite in the fault core; ii) pristine Macigno sandstone in an undeformed domain far from the main slip surface or any subsidiary shear plane.

Chemical analysis of the foliated cataclasite, compared to the pristine sandstone, shows (Table 5.1): i) marked K_2O increment; ii) slight CaO increment; iii) slight Na_2O , MgO and SiO_2 decrease. Combining such chemical variations and the mineralogical phases recognized in the preliminary XRD analysis, it is possible to estimate that the cataclasite is characterized by (Table 5.2): i) slight quartz and chlorite decrease; ii) slight calcite and illite increment; iii) marked albite decrease; iv) marked orthoclase increment.

Diffraction analysis on oriented slides of the $< 2 \mu\text{m}$ fraction indicates (Figure 5.12): i) the different pre-XRD treatments do not produce remarkable differences in the results; ii) ethylene glycol solvated samples, compared to air dried ones, do not present a migration toward the lower angles of the diffraction peaks; iii) the foliated cataclasite, compared to the pristine sandstone, is characterized by lower Illite and Chlorite crystallinity, as indicated by the higher FWHMs (Full Width at Half Maximum) of the respective basal diffraction peaks.

5.5.3 Fluid inclusion microthermometry

5.5.3.1 Site 1

At Site 1, fluid inclusion microthermometry was performed on quartz crystals from mixed veins in chert nodules from fault gouge, and on calcite veins in Maiolica limestone from the hangingwall damage zone. Thirty-three fluid inclusions, organized in 5 FIAs, were analyzed in quartz crystals from two different veins (samples VL304, VL305). All measured inclusions are biphasic at room temperature and small-sized ($2 - 3 \mu\text{m}$); larger inclusions are not present and smaller inclusions could not be used for analysis (Figure 5.13A). Homogenization temperatures from quartz are clustered in the $140 - 170^\circ\text{C}$ range (Figure 5.13D). First ice melting temperatures (T_{fm}) could

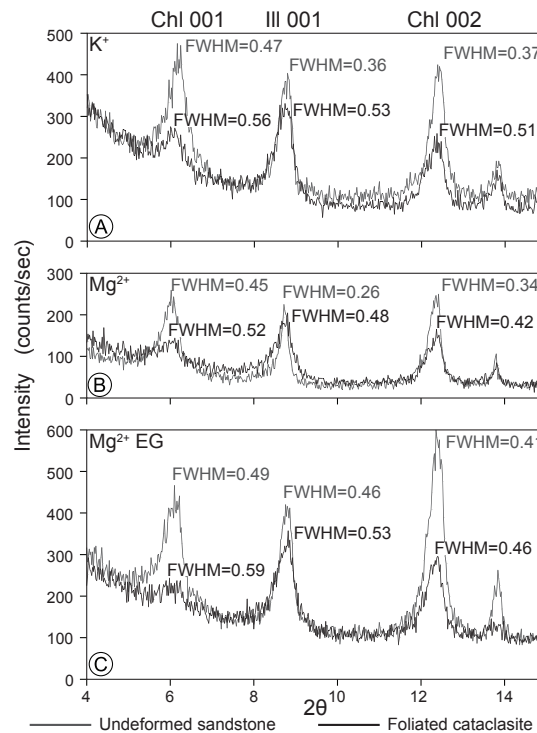
Table 5.1: Chemical composition of samples based on XRF analysis. VL₃₁₅: undeformed Macigno sandstone; VL₃₁₂: foliated cataclasite from Site 2 fault core.

	Na ₂ O	MgO	Al ₂ O ₃	SiO ₂	P ₂ O ₅	K ₂ O	CaO	TiO ₂	MnO	Fe ₂ O ₃	L.O.I.	sum
VL ₃₁₂	2.01	2.52	12.22	63.19	0.14	3.54	5.77	0.52	0.08	3.7	5.86	99.55
VL ₃₁₅	3.02	3.2	12.16	66.21	0.12	1.49	4.43	0.52	0.09	3.91	4.8	99.95

Table 5.2: Computed mineralogical compositions of samples based on chemical and XRD analyses (see text for details). VL₃₁₅: undeformed Macigno sandstone; VL₃₁₂: foliated cataclasite from Site 2 fault core.

	Quartz	Calcite	Albite	Anorthite	Orthose	Illite	Clorite
VL ₃₁₂	31.7	8	17.7	5.6	14.6	11.5	10.9
VL ₃₁₅	35.3	6	26.2	5.1	4.9	9	13.3

Figure 5.12: XRD of oriented slides of the < 2 μm grain-size fraction, subject to different treatments before analysis: A) K^+ saturated and air dried; B) Mg^{2+} saturated and air dried; C) Mg^{2+} saturated and ethylene glycol solvated (Mg^{2+} EG). Numbers indicate the Full Width at Half Maximum (FWHM) of the Chlorite (Chl) and Illite (Ill) basal diffraction peaks.



be measured only in few inclusion; values ranged from -20 to -15°C and are considered indicative of a H_2O - NaCl fluid system. Final ice melting temperatures ($T_{\text{m ice}}$) could be measured in only 9 inclusions, and values are in the range -2.7 to -0.9°C , indicative of a low salinity water ranging from 4.5 to 1.6 eq. wt % NaCl (Bodnar, 1993). Further evidence of a low salinity water is given by the metastable behavior of 12 more inclusions (Figure 5.13E).

Seven inclusions were analyzed in calcite veins from the hanging-wall damage zone. Secondary fluid inclusions aligned along the twinning planes were not taken into account, while isolated inclusions in the zones between the twinning planes were analyzed. All measured inclusions are biphasic at room temperature and small-sized (2 - 3 μm); larger inclusions are not present and smaller inclusions could not be used for analysis (Figure 5.13B). Homogenization temperatures range between $120 - 140^\circ\text{C}$ (Figure 5.13D). A $T_{\text{m ice}}$ temperature of -0.9°C could be measured in one inclusion, corresponding to a salinity value of 1.6 eq. wt % NaCl (Bodnar, 1993).

5.5.3.2 Site 2

Twenty-six fluid inclusions were analyzed in the S-C deformed calcite veins in the Maiolica limestone from the footwall damage zone, close to the master slip surface. Isolated inclusions and FIAs in the zones between the twinning planes were analyzed, while secondary fluid inclusions aligned along the twinning planes were not further considered. All measured inclusions are biphasic at room temperature and small-sized (2-3 μm); larger inclusions are not present and smaller in-

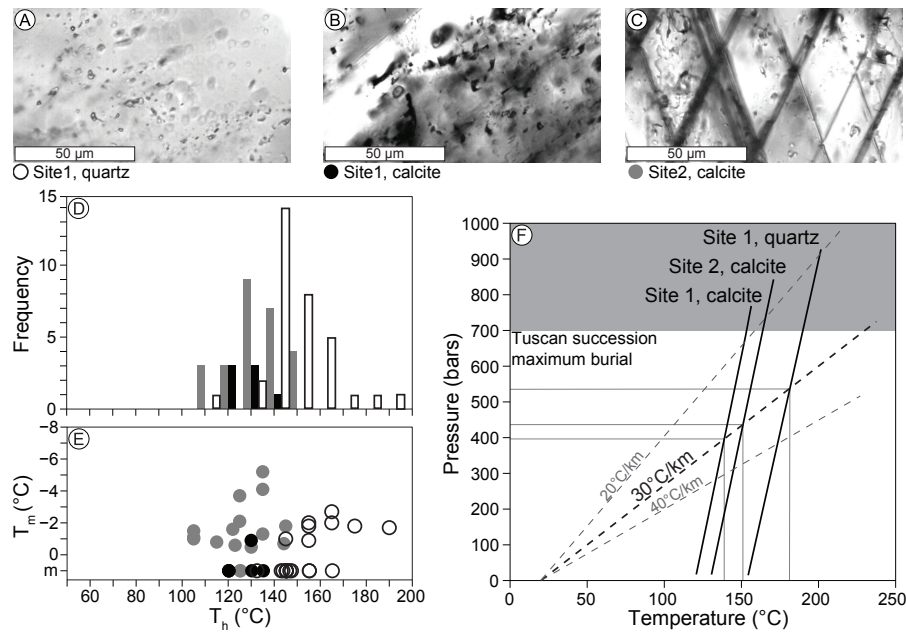


Figure 5.13: Fluid inclusion microthermometry analysis on different vein infillings (see text for details). A) Analyzed inclusions in quartz crystals from mixed veins at site 1 fault core. B) Analyzed inclusions in calcite veins at site 1 hangingwall damage zone. C) Analyzed inclusions in S-shaped veins at site 2 footwall damage zone. D) Homogenization temperature (T_h) frequency plots, bin size 10°C (bin colors as in A, B and C). E) Homogenization temperature (T_h) plotted versus final ice melting ($T_{m\ ice}$) temperature (bin colors as in A, B and C). F) Temperature-Pressure diagrams to compute real trapping conditions. Thick continuous lines: isochoric paths of specific samples. Dashed lines: hydrostatic gradients used as independent constrains. Shaded area indicates P-T conditions deeper than the maximum burial experienced by Tuscan succession (see text for details).

clusions could not be used for analysis (Figure 5.13C). Homogenization temperatures from calcite range between 110 and 150°C (Figure 5.13D). First ice melting temperatures (T_{fm}) could be measured only in few inclusion; values ranged from -20 to -15°C and are considered indicative of a H₂O-NaCl fluid system. $T_{m\ ice}$ temperatures could be measured in 15 inclusions and are between -5.2 and -0.2°C (Figure 5.13E), indicative of variable salinity between 8.1 and 0.33 eq. wt % NaCl (Bodnar, 1993). The majority of the inclusions (12) showed $T_{m\ ice}$ temperatures higher than -2.1°C, indicative of low salinity water with less than 3.5 eq. wt % NaCl.

5.5.4 Stable isotopes

Stable isotope analysis for oxygen and carbon was performed on carbonate host rocks, calcite veins and quartz veins from fault related and non fault related samples from sites 1, 2 and 3 (Figure 5.14A). In the relatively undeformed Site 3, calcite host rocks show $\delta^{18}\text{O}$ val-

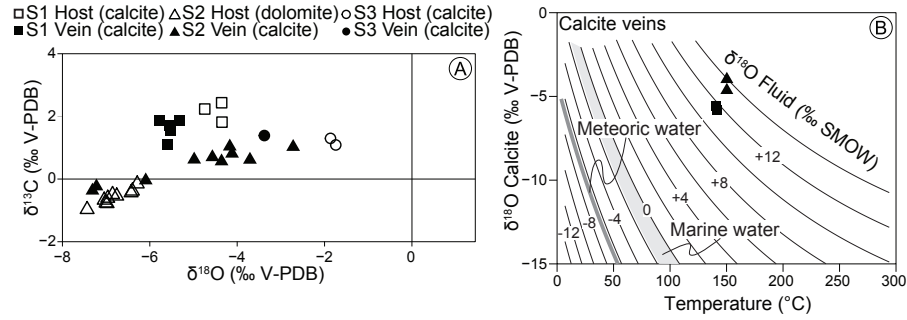


Figure 5.14: A) Oxygen and carbon stable isotopes from veins and host rocks of samples from Site 1 and Site 2 fault damage zones, and from the relatively undeformed Site 3. B) Oxygen isotope fractionation during equilibrium precipitation: calcite veins $\delta^{18}\text{O}$ and parent fluid composition curves as a function of temperature. Approximate isotopic composition of meteoric and marine water is highlighted in dark and light grey respectively (see text for details).

ues of $\sim -1.8\text{‰}$ V-PDB and $\delta^{13}\text{C}$ values of $\sim +1.1\text{‰}$ V-PDB. Calcite veins from the same outcrop show similar C isotope values and more depleted O isotope values of $\sim -3.4\text{‰}$ V-PDB. At Site 1, calcite host rock from the hangingwall damage zone show $\delta^{18}\text{O}$ values around -4.5‰ V-PDB and $\delta^{13}\text{C}$ values of $\sim +2\text{‰}$ V-PDB. Calcite veins from the same outcrop show slightly depleted C isotope values of $\sim +1.5$ and O isotope values around -5.5‰ V-PDB. At Site 2, the dolomitic host rock from footwall damage zone shows more depleted stable isotope values, with $\delta^{18}\text{O}$ ranging from -7.45 to -6.23‰ V-PDB and $\delta^{13}\text{C}$ ranging from -0.94 to -0.12‰ V-PDB. Most of the calcite veins from the same samples are generally more enriched, with $\delta^{18}\text{O}$ between -4.99 and -2.72‰ V-PDB and $\delta^{13}\text{C}$ between $+0.6$ and $+1.07\text{‰}$ V-PDB. Only three calcite veins differ from this pattern and are much more depleted, with values similar to the dolomitic host.

5.6 DISCUSSION

The lower illite crystallinity in the fault cataclasite, compared to the undeformed Macigno sandstone, indicate that the activity of the Val di Lima low-angle extensional fault system took place during the exhumation history of the Tuscan succession, after it underwent its maximum tectonic burial. Better constraints on the P-T conditions at which deformation took place, and on the fluid circulating during fault activity, come from the few mixed quartz-calcite veins in chert nodules from the Site 1 fault core. Indeed, quartz crystals in such veins do not show any evidence of intracrystalline deformation nor recrystallization, and microthermometry data are considered to be representative of P-T conditions during fault activity. The measured homogenization temperatures of fluid inclusions in quartz crystals are well clustered, with $155 \pm 7^\circ\text{C}$ as the most representative value,

corresponding to $\sim 180^{\circ}\text{C}$ and 5.2 km pressure-corrected trapping conditions, constrained by the specific isochore and the assumption of a $30^{\circ}\text{C}/\text{km}$ geothermal gradient (Figure 5.13F).

Limestones and calcite veins from both Site 1 and Site 2 damage zones show evidences of recrystallization and therefore no information on the original fluids circulating during fault activity is preserved. Indeed, calcite recrystallization is likely due to abundant fluid circulation in the fault zone during later deformation phases. Considering fluid inclusion microthermometry on calcite veins, with the same assumption made for Site 1 quartz crystals, it is possible to estimate that calcite recrystallization took place at $140^{\circ}\text{C} / 4 \text{ km}$ and $150^{\circ}\text{C} / 4.2 \text{ km}$, respectively for Site 1 and Site 2 (Figure 5.13F). The parent fluid isotopic signature, computed from vein trapping temperatures and the respective measured $\delta^{18}\text{O}$, is remarkably enriched, with values as high as +11.5 and +14‰ SMOW for Site 1 and Site 2, respectively (Figure 5.14B). The low salinity and enriched isotopic signature of the parent fluid can be compatible with different scenarios, including: i) surface-derived water that infiltrated into the shallow crust and was affected by abundant fluid-rock interactions during its downward migration; ii) residual fluids expelled from the Macigno sandstone during due to tectonic compression; iii) dehydration of a shale sequence such as the Scaglia Formation. The dolomitic host rock in Site 2 is also affected by recrystallization, as clearly seen under cathodoluminescence microscopy, but in this case sample staining suggests no iron enrichment. The remarkably more depleted isotopic signature and the lower abundance of iron, suggest that dolomite recrystallization took place at different conditions than calcite recrystallization.

Within this framework, the Val di Lima fault system provides an interesting example of a low-angle extensional fault zone developed at shallow-crustal depths, showing important variations of deformation features and fluid-rock interactions during deformation. These variations are greatly influenced by the different lithologies involved in the fault system. Site 1 and Site 2, in particular, show development of different deformation structures in the Maiolica carbonates, which are related to their structural position with respect to the main slip surface (hangingwall versus footwall) and to the local fluid pressure. Site 2 shows evidences of fluid overpressure, as suggested by: i) the abundance of calcite veins in Maiolica carbonates in the footwall damage zone, increasing up to extreme values in the proximity of the main slip surface; ii) the contrasting almost complete absence of veining in the hangingwall damage zone; iii) the presence of a well-developed foliated cataclasite in the fault core, which could have acted as an efficient hydraulic barrier. These peculiar hydraulic conditions favored weakening and shear localization in Maiolica carbonates, as schematized in figure 5.15. In early deformation phases, Maiolica carbonates deformed by veining and pressure dissolution,

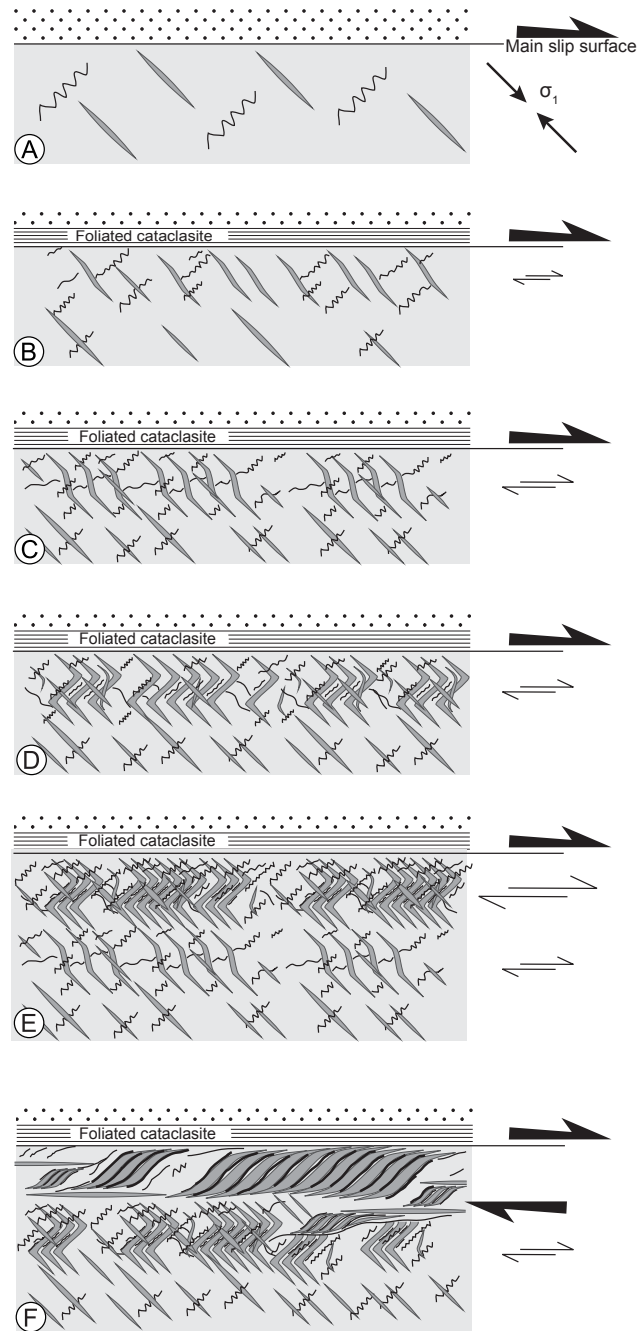


Figure 5.15: Evolution of the deformation fabric in the Maiolica limestone at site 2 footwall damage zone. Dotted: Macigno sandstone; light gray: Maiolica carbonates; dark gray: calcite veins; black: pressure-solution seams. A to F indicate successive evolution phases (see text for details).

leading to structures respectively oriented parallel and perpendicular to the maximum compression axis σ_1 (Figure 5.15A). Subsequently, vein formation increased in abundance, possibly favored by the combined conduit-barrier behavior of the fault system, and started to be arranged in fault-parallel “en echelon” arrays (Figure 5.15B). With progressive deformation, the vein arrays acted as semi-brittle shear zones and the typical tension-gashes fabric developed by propagation of the vein tips along the σ_1 direction and contemporaneous passive rotation of the veins’ central portions. Similarly, old pressure solution seams were rotated and evolved into slickolites (Figure 5.15C and D). Contemporaneously, new veins and stylolites formed, still oriented parallel and perpendicular to the σ_1 axis, respectively (Figure 5.15C and D). Subsequently, due to further pressure dissolution and veining, the fault rock became almost completely vein dominated, with only few small preserved host fragments (Figure 5.15E). The extremely high vein/host rock ratio and the favorable orientation of the rotated veins, lead to remarkable strain-softening and shear localization in the central portion of the vein arrays. Eventually, the external portions of the arrays got dissolved, and a “ductile-like” shear zone with a characteristic S-C shear fabric formed by: i) deformation of the pre-existing structures (veins and slickolites); ii) opening of new fault-parallel veins (Figure 5.15F). As a consequence of this process, deformation patterns with opposite asymmetries can be presently observed within a single fault zone, characterized by a consistent kinematic. The Z-shaped tension gash fabric is still preserved in the outer damage zone, while the S-C shear fabric characterizes the inner damage zone, in the proximity of the main slip surface (Figures 5.7B and 5.15F).

The geological framework in which the Val di Lima low-angle fault system formed is schematized in figure 5.16. i) Early contraction leads to development of a thrust-related anticline in the Tuscan succession. The thrust fault has a ramp-and-flat geometry characterized by a long flat segment, located along the Triassic evaporites at the base of the succession. During this phase the fault zone is dominated by circulation of low-saline warm fluids (Figure 5.16A). ii) Further contraction leads to fold amplification in the Mesozoic succession and development of an overturned fold by progressive hinge migration (e.g., Suppe & Medwedeff, 1990; Storti & Salvini, 1996; Perrin et al., 2013). The fold has an isoclinal geometry, with the axial plane localized in the Triassic evaporites. The Macigno Formation, located at the top of the Tuscan succession, is not involved in the overturned fold and accommodates shortening by roof duplexing and thrust splaying (Figure 5.16B). This inference is supported by subsurface data in the seismic profile of figure 3, showing that the thickness of the overturned limb is appropriate for the involvement of the carbonate dominated multilayer, but by far too small for including the Macigno sandstone as well. Accordingly, we propose that the Macigno detaches from the

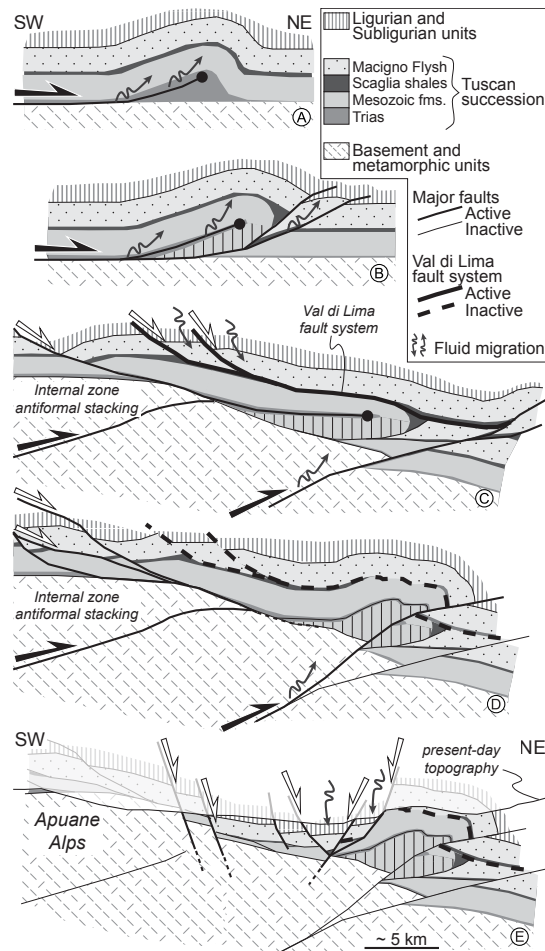


Figure 5.16: Cartoon illustrating the proposed structural evolution of the Val di Lima anticline and detachment. A to E indicate successive tectonic phases (see text for details).

underlying carbonates along the Scaglia Formation shales. iii) Uplift in the rear Apenninic thrust wedge, related to deep-rooted thrusting and antiform stacking, leads to the passive rotation and foreland-ward tilting of the isoclinal fold. Within this contractional regional framework, the increased wedge taper above the critical values (Davis et al., 1983) triggers extensional deformations in the upper part of the thrust wedge. The Val di Lima low-angle fault system develops at this stage and provides a relatively long flat-segment of a regionally-sized detachment, localized in the Scaglia shales. The permeability increase associated with the extensional fault systems allows fluid migration through the crust, and the low salinity values indicate that such fluids did not interacted with the Triassic evaporites at the base of the Tuscan succession (Figure 5.16C). iv) When contraction resumes in this sector of the Apenninic thrust wedge, the activity of a more internal deep-rooted thrust splay affects the Val di Lima isoclinal fold, partially crosscutting its reversed limb. Associated hangingwall deformations cause re-folding of the isoclinal fold limbs and the Val di Lima

fault system, which is now inactive (Figure 5.16D). v) Development of large-scale high-angle extensional fault systems crosscutting the Val di Lima detachment and surface erosion produce the present-day topography (Figure 5.16E). Direct dating of the Val di Lima low-angle extensional fault system is not available, but the timing of fault activity can be constrained from the geological framework of the region. Indeed, the early compression (Figure 5.16A and B) affects the Lower Miocene foredeep deposits of the Macigno and Cervarola units (Chicchi & Plesi, 1991, 1992; Plesi et al., 2002). The late orogenic high-angle faulting (Figure 5.16E) is associated with the opening of the Garfagnana basin, in which continental deposits as old as Lower Pliocene are present (Bartolini & Bortolotti, 1971; Calistri, 1974; Dallan et al., 1981; Nardi et al., 1986, 1987; Fazzuoli et al., 1998). Therefore, the activity of the low-angle extensional fault system (Figure 5.16C) and its successive refolding due to out-of-sequence thrusting (Figure 5.16D) can be reasonably constrained to the upper Miocene.

5.7 CONCLUSION

The structural architecture exposed in the Lima valley is characterized by mutually overprinting contractional and extensional deformation structures, developed during the long-term evolution of the Northern Apenninic thrust wedge. The Val di Lima low-angle fault system, in particular, affected the Tuscan non-metamorphic unit during its progressive exhumation and indicates occurrence of extensional stages within the orogenic contractional evolution of the thrust wedge.

The Val di Lima low-angle detachment is considered the flat portion of a staircase extensional fault system, developed in the shallower portion of the Apenninic thrust wedge to compensate for its overcritical taper, produced by antiformal stacking of the internal zone due to deep-rooted thrusting. The Val di Lima detachment acted as a combined conduit-barrier system which affected fluid circulation within the upper crust in two ways: i) allowed the migration of low-salinity fluids, due to the increased permeability along the fault zone; ii) favored footwall fluid overpressures, localized in the places where an impermeable lithology (e.g. foliated cataclasite) developed in the fault core and acted as an efficient hydraulic barrier. Abundant fluid circulation in fault damage zones also characterized the late-stage evolution of the low-angle fault system, allowing the recrystallization of calcite veins and limestone host rocks at shallower conditions (~ 4 km).

These observations have implications with regard to our understanding of deformation of the upper crust. Important stages of wedge extension, similar to the one described here, are likely to occur during the geological history of most thrust wedges, because their long-term evolution is characterized by complex interactions among tectonics,

gravitational body forces, and surface processes. The systems of brittle extensional fault zones, resulting from such extensional pulses, affect fluid circulation through the upper crust by producing articulated networks of hydraulic conduits, barriers, or mixed conduit-barrier systems. In particular, as demonstrated by our results, the effects of extensional fault zones on fluid circulation is twofold: i) they provide effective pathways allowing fluid migration; ii) they can trigger fluid overpressuring, especially in the footwall of shallow-dipping fault segments. Eventually, fluid circulation can exert a strong influence on the mechanical behavior of thrust wedges either by reducing the effective normal stress at depth or triggering the formation of hydrous clay minerals lowering the frictional properties of fault zones.

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TELLARO DETACHMENT

This chapter is presented in the form of a manuscript that is currently been prepared for submissions to the *Geological Society of America Bulletin*. It describes the structural architecture and petrographical, geochemical, and microthermometric features of the Tellaro large scale low-angle extensional fault system, developed in the internal portion of the Northern Apennines thrust wedge.

SUBMITTED ARTICLE

Structural anatomy and paleofluid evolution of the Tellaro detachment fault system, Northern Apennines, Italy.

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STRUCTURAL ANATOMY AND PALEOFLUID EVOLUTION OF THE TELLARO DETACHMENT FAULT SYSTEM, NORTHERN APENNINES, ITALY.

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ABSTRACT

Late Miocene low-angle and Plio-Pleistocene high-angle extensional faults occur in the internal part of the northern Apennines thrust wedge. Among the former, the Tellaro Detachment is very well exposed, cut through carbonate rocks within the brittle crust, and is responsible for remarkable thinning of the non-metamorphic Tuscan succession. The structural architecture of the Tellaro Detachment is investigated through detailed structural mapping, restoration of the superimposed deformations, and damage zone characterization. Pressure-depth conditions and palaeofluid evolution of the fault system are studied through microstructural, petrographic, geochemical and fluid inclusion analysis of faulted rocks and fault-related dolomite and calcite veins.

Our results indicate that fluid circulation during the long-lasting activity of the Tellaro Detachment caused precipitation of syn-kinematic dolomite and calcite vein infillings at gradually decreasing depths (~ 5 to 2 km) and temperatures (~ 175 to 90°C). During incipient faulting, the system was hydraulically closed and dominated by high-salinity fluids. Successively, large-scale extensional fault zones allowed mixing of low- and high-salinity fluids, in an open hydraulic system. Finally, when the fault system was almost completely exhumed and became inactive, a later shallow circulation system (~ < 1.5 km) became dominated by meteoric fluids.

We conclude that the Tellaro Detachment, developed during the progressive exhumation of the internal portion of the Northern Apennines thrust wedge, was characterized by: i) contemporaneous activity of low-angle master faults and subsidiary high-angle faults; ii) elevated fluid pressures and low differential stress conditions. Our findings support the involvement of misoriented faults in continental crustal thinning.

6.1 INTRODUCTION

Understanding whether extensional detachment fault systems can be activated at very shallow dip, or result from the passive rotation of former high-angle “Andersonian” fault zones, is a matter of long-lasting debate involving field investigation, geochemical and geophysical evidence, analogue and numerical modeling (e.g., Yin, 1989; Buck, 1991; Axen, 1992; Lavier et al., 1999; Lavier & Buck, 2002; Ranero & Pérez-Gussinyé, 2010; Collettini, 2011). Despite such an impressive amount of research, the conundrum is not yet resolved. The question whether slip can occur along mechanically misoriented slip surfaces, allowing the routinely coexistence of low- and high-angle faults, is still open. The Apennines fold-and-thrust belt, in the Italian peninsula, is a good place to investigate such topic, since many exhumed or active low-angle extensional fault systems have been documented in recent years and associated to syn- or post-orogenic crustal extension regimes (e.g., Elter et al., 1975; Carmignani & Kligfield, 1990; Carmignani et al., 1994, 1995; Lavecchia et al., 1994; Bartole, 1995; Storti, 1995; Barchi et al., 1998; Boncio et al., 2000; Collettini & Barchi, 2002; Argnani et al., 2003; Brogi, 2004, 2008; Collettini & Holdsworth, 2004; Collettini et al., 2006b; Chiaraluce et al., 2007; Hreinsdottir & Bennett, 2009).

Fluid circulation has been demonstrated to play a fundamental role to favor the activity of low-angle extensional fault systems, due to the typical conduit-barrier hydraulic architecture of brittle shear zones in cohesive low-porosity rocks (e.g., Caine et al., 1996; Storti et al., 2003), which favors fluid overpressuring (Gudmundsson, 2001; Miller et al., 2004; Collettini et al., 2006a) and the formation of low-friction hydrous clay minerals (Hubbert & Rubey, 1959; Davis et al., 1983; Dahlen et al., 1984; Sibson et al., 1988; Dahlen, 1990; Moore & Vrolijk, 1992; Kirschner & Kennedy, 2001; Collettini & Holdsworth, 2004; Fagereng et al., 2010; Lacroix et al., 2013). This can occur both when deep hot fluids migrate upward, or cold surface-derived fluids infiltrate at depth (Wickham & Taylor, 1987; Fricke et al., 1992; Morrison, 1994; Morrison & Anderson, 1998; Bebout et al., 2001; Mulch et al., 2004, 2006; Verhaert et al., 2004, 2009; Person et al., 2007; Ingebritsen & Manning, 2010; Rossetti et al., 2010; Gébelin et al., 2011; Gottardi et al., 2011).

In this paper, we describe the structural architecture and the paleofluid evolution of the Tellaro Detachment (Storti, 1995), a low-angle extensional fault system outcropping in the innermost portion of the Northern Apennines (Italy), which was active during the late Cenozoic exhumation of the axial zone of the mountain belt. Our data indicate that anastomosed near-horizontal fault zones, and high-angle subsidiary faults in the hanging-wall, were contemporaneously active in a fluid-assisted environment. Results presented in this study con-

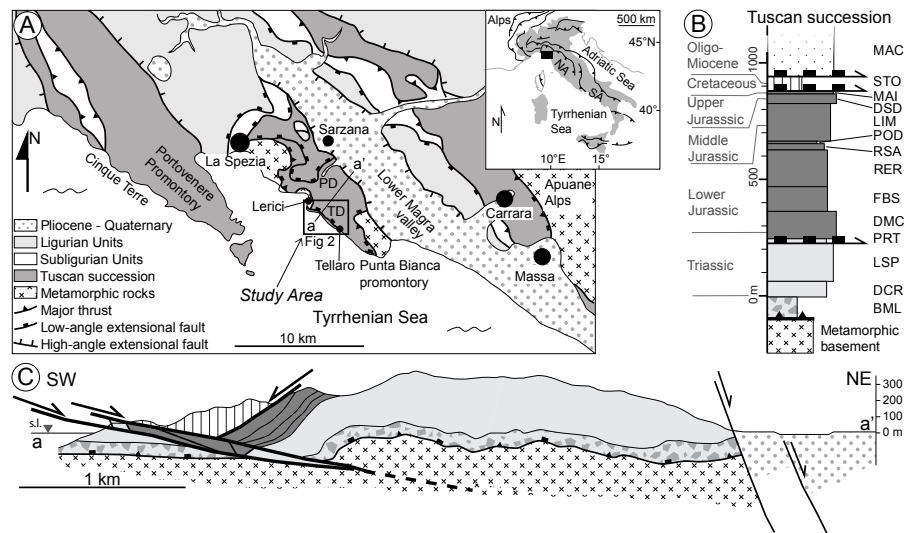


Figure 6.1: A) Tectonic map of the Punta Bianca promontory and the surrounding region (location of the study area is in the inset). NA: Northern Apennines; SA: Southern Apennines; PD: Pitelli Detachment extensional fault system; TD: Tellaro Detachment extensional fault system. B) Stratigraphic column of the Tuscan succession in the Punta Bianca promontory area (after Servizio Geologico d'Italia, 2005) with indicated the position of the main extensional detachments. MAC: Macigno Formation; STO: Scaglia Toscana Formation; MAI: Maiolica Formation; DSD: Diaspri Formation; POD: Marne a Posidonia Formation; LIM: Calcare Selcifero di Limano Formation; RSA: Rosso Ammonitico Formation; RER: Ferrera Formation; FBS: Biassa Formation; DMC: Dolomie del Monte Castellana Formation; PRT: Portoro Formation; LSP: La Spezia Formation; DCR: Dolomie di Coregna Formation; BML: Breccie di Maralunga Formation. C) Geological cross-section through the Punta Bianca promontory (modified from Federici & Raggi, 1975; Servizio Geologico d'Italia, 2005). Same colors as B; thick black lines indicate the Tellaro Detachment and associated structures.

tribute to support the notion that continental crustal thinning and fault activity can occur also for shallow values of fault dip.

6.2 GEOLOGICAL SETTING

The Apenninic segment of the Alpine-Himalayan orogenic system has been growing on top of the retreating subduction of the Adria plate below the European plate in Cenozoic times (Malinverno & Ryan, 1986; Royden, 1988; Dewey et al., 1989; Patacca et al., 1990; Doglioni, 1991). The remnants of the Jurassic Ligurian ocean and of the thinned passive margin of the Adria Plate are accreted in the Northern Apennines (Boccaletti et al., 1971; Elter & Pertusati, 1973), resulting in a complex three-dimensional orogenic architecture (Figure 6.1A; Elter, 1975). In its internal (westernmost) portion, the thrust wedge includes, from bottom to top: (i) the Apuane and Massa metamorphic units, reaching, HP greenschist facies (estimated deepest

burial ~ 20 and 25 km, respectively; Di Pisa et al., 1985; Franceschelli et al., 1986; Molli et al., 2000a,b, 2002); (ii) the non-metamorphic Triassic to Lower Miocene basinal succession of the “Tuscan Nappe” (Figure 6.1B); (iii) the non-metamorphic Jurassic-Cretaceous Subligurian and Ligurian ophiolite-bearing units and the Middle Eocene to Pliocene Epiligurian wedge-top sediments (Elter, 1975; Molli, 2008). The first-order nappe pile underwent significant tectonic thinning by low-angle extensional faulting (Carmignani & Kligfield, 1990), then overprinted by out-of-sequence thrusting (Storti, 1995; Argnani, 2002; Molli, 2008; Boccaletti et al., 2011) and by further extensional deformations that also triggered Upper Miocene to Pleistocene magmatic activity in the southern portion of the belt (Elter et al., 1975; Carmignani et al., 1995; Storti, 1995; Barchi et al., 1998; Decandia et al., 1998; Jolivet et al., 1998; Brogi, 2004, 2008; Collettini et al., 2006b; Rossetti et al., 2008). The deformation conditions of the Tuscan basinal succession within the Apenninic thrust wedge have been constrained through different methodologies including vitrinite reflectance (Reutter et al., 1983), stable isotopes (Carter & Dworkin, 1990), fission track on apatite (Abbate et al., 1994; Balestrieri et al., 2003; Zattin et al., 2002) and zircon (Bernet et al., 2004; Fellin et al., 2007), fluid inclusion microthermometry (Montomoli et al., 2001; Montomoli, 2002), illite crystallinity (Cerrina Feroni et al., 1983; Carosi et al., 2003), calcite-dolomite geothermometry (Carosi et al., 2003) and TEM microstructure analysis (Molli et al., 2011). The maximum tectonic burial of the Tuscan succession, related to the first thrusting event, took place at about $15 - 10$ Ma, with estimated conditions of ~ 7 km depth and $200 - 250^{\circ}\text{C}$ in the internal portion of the Northern Apennines (La Spezia area; Molli et al., 2011), and progressively shallower conditions towards the east (e.g., Fellin et al., 2007, and references therein). The nominal geothermal gradients corresponding to the different paleotemperatures calculated by the different authors range between 18 and $41^{\circ}\text{C}/\text{km}$ (Molli et al., 2011, and references therein) with an average value of $31 \pm 4^{\circ}\text{C}/\text{km}$.

The Punta Bianca promontory, in the innermost part of the Northern Apennines, exposes the complete stack of the nappe pile, folded in a northwestward plunging anticline (Figure 6.1). Deformation in this area is polyphasic and includes (Figure 6.1C): i) the NE-vergent stacking of first order tectonostratigraphic units during early-middle Miocene; ii) middle to late Miocene antiformal stacking and duplexing of previously stacked units and related gravity-driven tectonic thinning mainly by low-angle normal faulting; iii) Pliocene to Quaternary crustal extension associated with the eastward migration of both the contractional deformation front and Tyrrhenian back-arc shoulder (Elter et al., 1975). Within this framework, the previously stacked and tectonically thinned units are affected by late-orogenic high-angle extensional and transtensional fault systems, associated with the present-day morphology. The Tellaro Detachment, investi-

gated in detail in this paper, is the best exposed among the late-orogenic low-angle extensional fault systems and is characterized by spectacular outcrops along the shoreline between the villages of Lerici and Tellaro (Figure 6.2). This fault system is mainly affecting and dismembering the km-thick Tuscan basinal succession, constituted by Triassic to Lower Cretaceous carbonates, Lower Cretaceous to Upper Oligocene shales, and Upper Oligocene to Lower Miocene flysch sandstones (Figure 6.1B). In its northernmost portion, nearby Lerici, the detachment crosscuts the basal tectonic contact between unmetamorphic upper Triassic limestones and the underlying Mid-Triassic metamorphic clastic rocks.

6.3 METHODOLOGY

Detailed mapping and structural analysis were carried out in the study area to unravel the three-dimensional architecture of the low-angle extensional fault system and its tectonic transport direction, and to characterize the associated damage zone. In each field station, tectonic transport direction was obtained by using the attitude of S-C fabrics, fault slickenlines and corrugations, and fault-related veins (Passchier & Trouw, 2005). Structural data are reported using the right-hand rule and stereographic projections are plotted on a lower hemisphere diagrams using the Daisy3 software (Salvini et al., 1999).

To characterize the fault rocks, the meso/micro-scale damage zone architecture, and the role of the fluids involved in the deformation, 42 oriented samples were collected along the exposed fault zones. For comparison purposes, additional sampling was carried out (i) along the Pitelli Detachment (Storti, 1995), another low-angle extensional fault zone geometrically overlying the Tellaro Detachment, and (ii) in a relatively undeformed area, several hundreds of meters away from major fault zones. In text and figures, samples are named with the PB label followed by a progressive 2-digit number. Since many veins are generally present in each sample and show remarkably different features, a suffix is added to refer to each of them (e.g., PB22-1, PB22-2). Each rock sample was cut to obtain two “mirrored” slabs, one of which was polished and stained with Alizarin red and potassium ferricyanide to easily discriminate different carbonate minerals such as calcite, dolomite and their respective iron-rich phases. From the other slab, thin-sections and wafers for microthermometry were prepared. Microstructural and petrographic analyses of 43 thin sections were carried out using optical and cold cathodoluminescence (CL) microscopy. The latter was performed with Techosyn model 8200 Mark II cold cathodoluminescence stage operated at 12 kV, 100-200 μ A gun current.

Stable isotope analysis was done on microsamples of veins and host rocks. Microsamples were drilled from the stained slabs (after

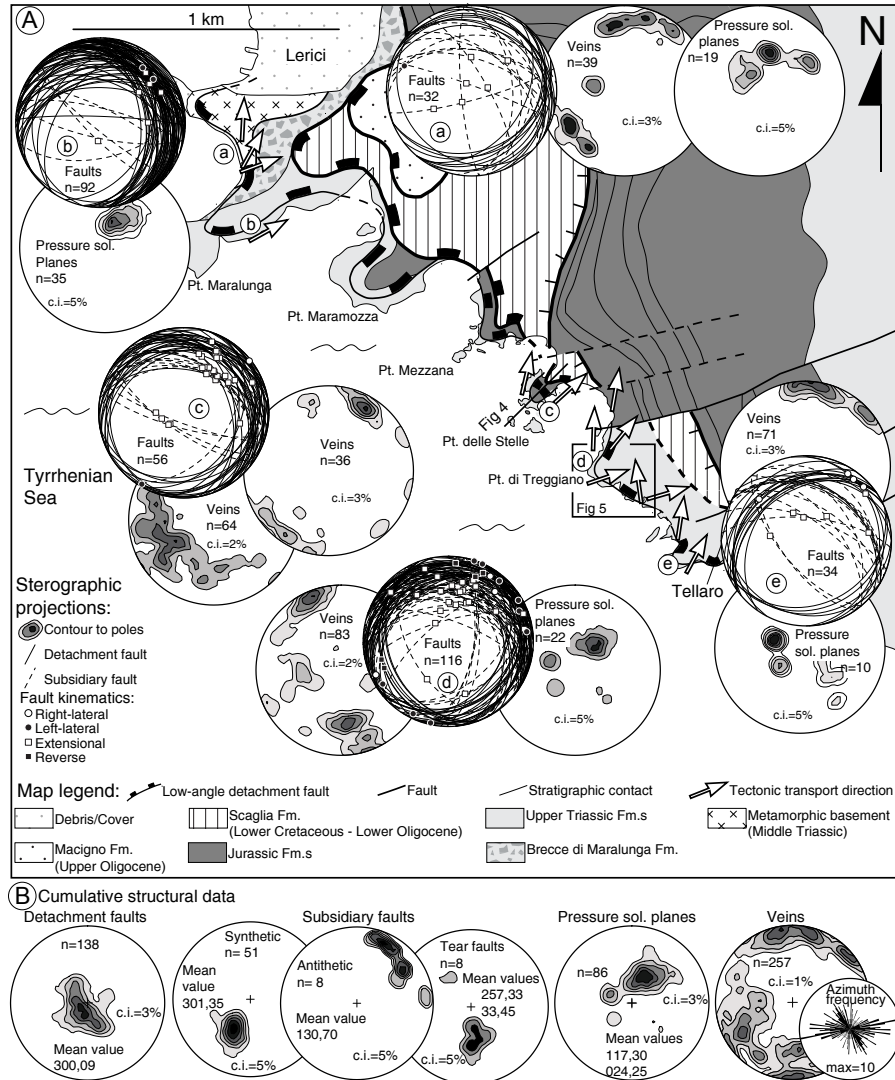


Figure 6.2: Structural map of the Tellaro Detachment (see Figure 6.1A for location). A: Geological map with the trace of the low-angle fault envelope, lower-hemisphere projections of structural data in the different sectors, and inferred tectonic transport directions indicated by white arrows (after Servizio Geologico d'Italia, 2005). B: Cumulative structural data for the whole exposed area of the Tellaro Detachment.

removal of the material altered by staining) with a dental drill. Carbonate powders were reacted with 100% phosphoric acid at 70°C using a Gasbench II connected to a Thermo Finnigan Delta V+ mass spectrometer. All values are reported in per mil (‰) relative to the Vienna PDB standard. Reproducibility was checked by replicate analysis of selected samples and laboratory standards and is better than ± 0.07 and ± 0.09 ‰ for oxygen and carbon isotope composition respectively. Isotope fractionation curves were computed with the formula

$$10^3 \ln \alpha = \frac{A 10^6}{T^2} + B \quad (6.1)$$

using $A = 2.78$, $B = -2.89$ (Friedman & O'Neil, 1977) and $A = 3.14$, $B = -2$ (Land, 1983) for calcite and dolomite, respectively.

Fluid inclusions were analyzed in doubly polished thick sections (wafers) which were mapped with a standard petrographic microscope. The wafers were subsequently broken into smaller pieces (chips). Whenever possible, the fluid inclusion assemblage (FIA) approach was followed (Goldstein & Reynolds, 1994). These fluid inclusion assemblages do not show any relationship to fractures, occur randomly in the sample and are considered to represent the trapping or recrystallization conditions of the minerals analyzed. Each FIA is composed of few inclusions (generally 5 to 10) considered to be representative of the same trapping event. A few isolated inclusions were measured as well, but only if many of them were found in the same vein. Samples with only a few isolated inclusions (less than 5 per vein) were not further considered. A Linkam THMSG600 heating-cooling stage was used for microthermometric analyses and calibrated weekly with SynFlinC synthetic standards. In order to avoid any artificial stretching of the inclusions, heating was always performed before freezing and each chip was heated only once. Individual chips were then examined through CL microscopy to exclude recrystallization of the crystals hosting the measured inclusions. Analyzed inclusions are 2 to 5 μm in size and monophasic (liquid) or two-phase (aqueous liquid + vapor bubble) at room temperature. Vapor bubble size ranges from 10 to 20% in volume in analyzed two-phase inclusions; inclusions with bigger bubbles are rare and were not further considered. Three temperatures were routinely measured in two-phase inclusions: homogenization to the liquid phase temperature (T_h), indicative of the minimum trapping temperature of the inclusion; first ice melting temperature (T_{fm}), indicative of the fluid system; final ice melting temperature ($T_{m\text{ ice}}$), inversely proportional to the salt content of the aqueous phase. Fluid salinity was computed from measured $T_{m\text{ ice}}$ values using the software FLUIDS (Bakker, 2003). Pressure correction was applied to the T_h values to determine the actual trapping conditions (temperature and pressure) of each FIA. The trapping conditions were defined by the intersection of the fluid pressure gradient (computed from the estimated geothermal gradient) and the appro-

priate isochore (computed from the salinity and T_h of each group of inclusions). Monophase inclusions were cooled to -20°C for two weeks in order to check their possible metastability. If they were still monophase after this treatment, a vapor bubble was artificially created via rapid heating/freezing cycles, and then T_{fm} and $T_{m\ ice}$ temperatures were measured in standard freezing experiments. Some inclusions showed a metastable behavior during re-heating after being frozen: their bubble returned to the original pre-freezing size at temperatures above 0°C despite the slow heating rate ($0.1^{\circ}\text{C}/\text{minute}$).

Strontium isotope analysis on calcite veins was performed on powders drilled from rock slabs with a manual driller equipped with a tiny dentist bit. Samples were leached in 1M ammonium acetate and digested in 2.5M HCl. Sr was separated by conventional cation exchange chromatography and total procedure blanks were less than 100 pg. $^{87}\text{Sr}/^{86}\text{Sr}$ was measured on a VG Sector 54-30 mass spectrometer operated in dynamic mode. Mass fractionation was corrected assuming $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ using an exponential law. NIST SRM987 gave 0.710249 ± 16 (2 SD, $n=17$).

6.4 RESULTS

6.4.1 *Fault system structural architecture*

The Tellaro extensional detachment includes low-angle master faults on which high-angle subsidiary faults sole down. The tectonic transport direction, measured in 14 different field sites, is toward the NE, with local values ranging from $\text{N}010\text{E}$ to $\text{N}080\text{E}$ (Figure 6.2). The degree of tectonic thinning is quite variable along strike. South of Lerici, the Upper Cretaceous shales of the Scaglia Formation (STO) are tectonically juxtaposed next to Upper Triassic limestones of the La Spezia Formation (LSP). To the southeast, near Punta delle Stelle, the Tuscan succession is reduced down to few tens of meters because Lower Cretaceous and Jurassic formations are either totally missing or preserved only as meter- to decameter-scale lenses bounded by anastomosed low-angle shear zones (Figure 6.3).

Low-angle master slip surfaces are generally organized in few juxtaposed segments with similar attitude and metric spacing. In most cases, the uppermost segment is characterized by the largest stratigraphic offset (Figure 6.2). The attitude of low-angle fault zones is clustered around a NW-SE strike and a shallow NE dip ($\text{N}300,09$; Figure 6.2B). The variability around the mean value is aligned along two orthogonal directions, almost parallel to the fault strike and dip direction, respectively. Subsidiary high-angle faults are abundant and confined in the footwall damage zones of the master low-angle fault segments; they sole down onto secondary low-angle faults, and strike either near parallel or near perpendicular to the local tectonic transport directions (Figure 6.4). Synthetic subsidiary faults, dipping to-

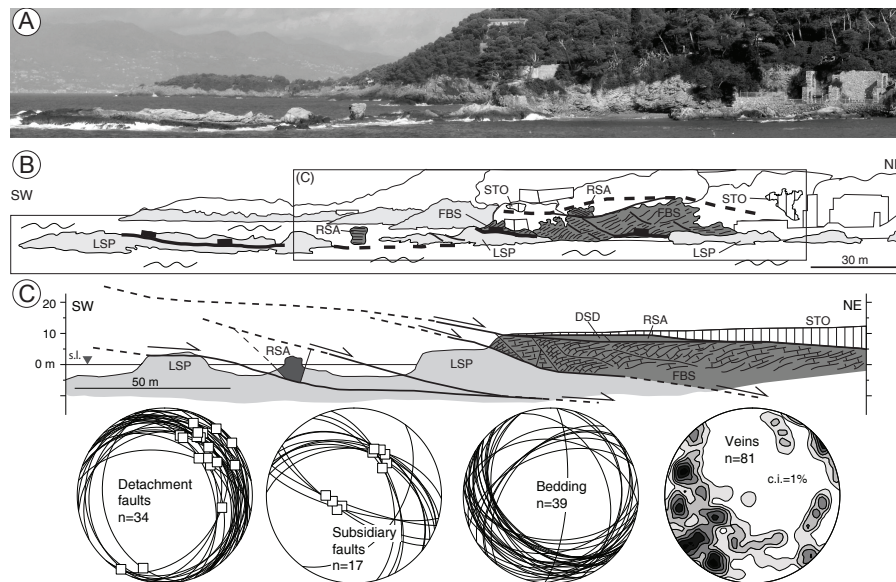


Figure 6.3: Panoramic view (A), interpretative line drawing (B) and schematic cross-section (C) of the Punta delle Stelle exposure (see Figure 6.2 for location). The Upper Cretaceous shales of the Scaglia Formation (STO) is tectonically juxtaposed onto anastomosed, metric- to decametric-scale, shear lenses of Jurassic rocks, which in turn overly the Upper Triassic limestones of the La Spezia Formation (LSP) (See Figure 6.1B for the pristine thickness of the succession and the abbreviations used). Structural data collected at this site are also provided.

ward the NE, are by far the most abundant among the subsidiary faults; they have extensional kinematics and moderate dip values. Antithetic subsidiary faults, dipping toward the SW, have extensional kinematics and steeper dip values. Subsidiary tear faults, striking parallel to the tectonic transport direction, have strike-slip kinematics and moderate dip values. Fault surfaces are generally marked by slickenlines and are characterized by centimeter- to meter-scale corrugations oriented parallel to the tectonic transport direction. In proximity of the biggest corrugations, the hangingwall damage zone is characterized by the presence of ductile deformation structures, such as tight folds with corrugation-parallel axes (Figure 6.4). Pressure-resolution S-C arrays are abundant along many low-angle master slip surfaces (Figure 6.5A). Their orientation is consistent with the kinematics obtained from slickensides, with a $N115,30$ mean attitude and pole distribution variability aligned along the NE-SW direction (Figure 6.2B). In places, dolomitic breccia bodies are associated with the low-angle fault zones. They have a tabular shape (1-2 m in height and several meters in length) with irregular boundaries, and are discordant with respect to bedding in the host formations (Figure 6.6). High-angle synthetic subsidiary faults systematically cut through the dolomitic breccias.

Subvertical syntectonic veins, filled by dolomite or calcite, are very abundant in the whole area, both in hangingwall and footwall dam-

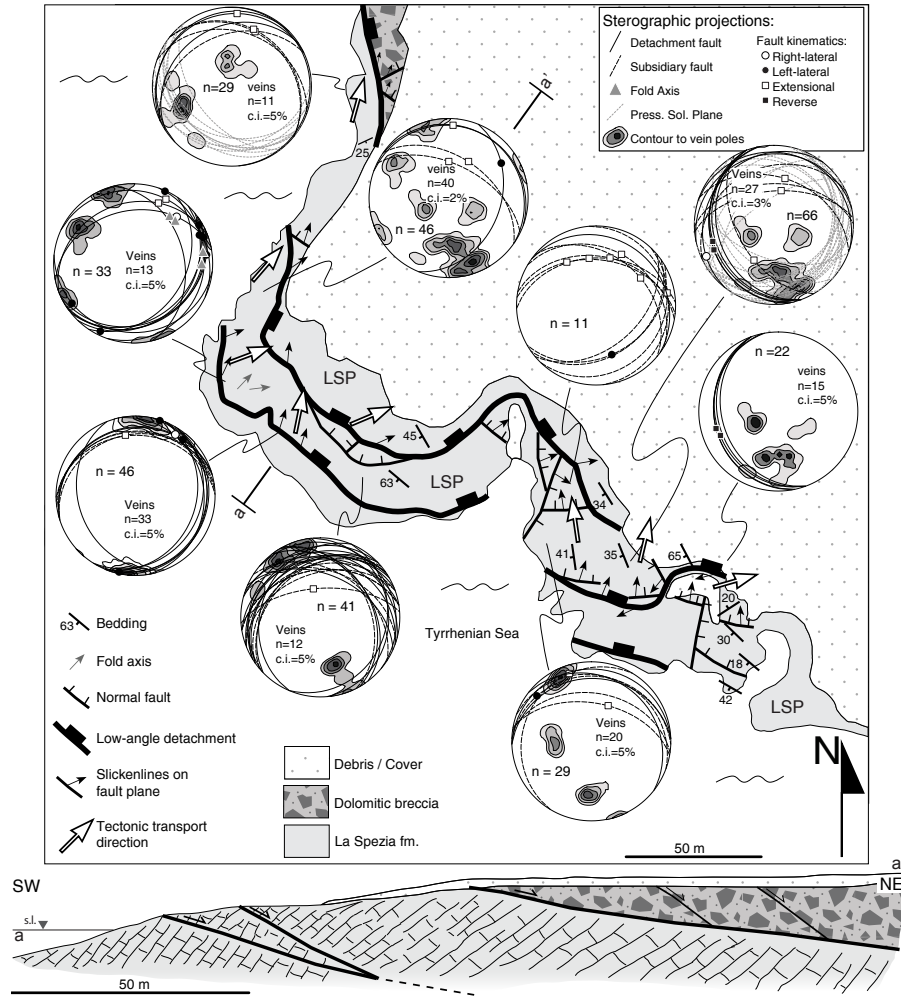


Figure 6.4: Detailed structural map of the Punta di Treggiano segment of the Tellaro Detachment (see Figure 6.2 for location), and corresponding geologic cross-section. Structural data collected in this sector are provided.

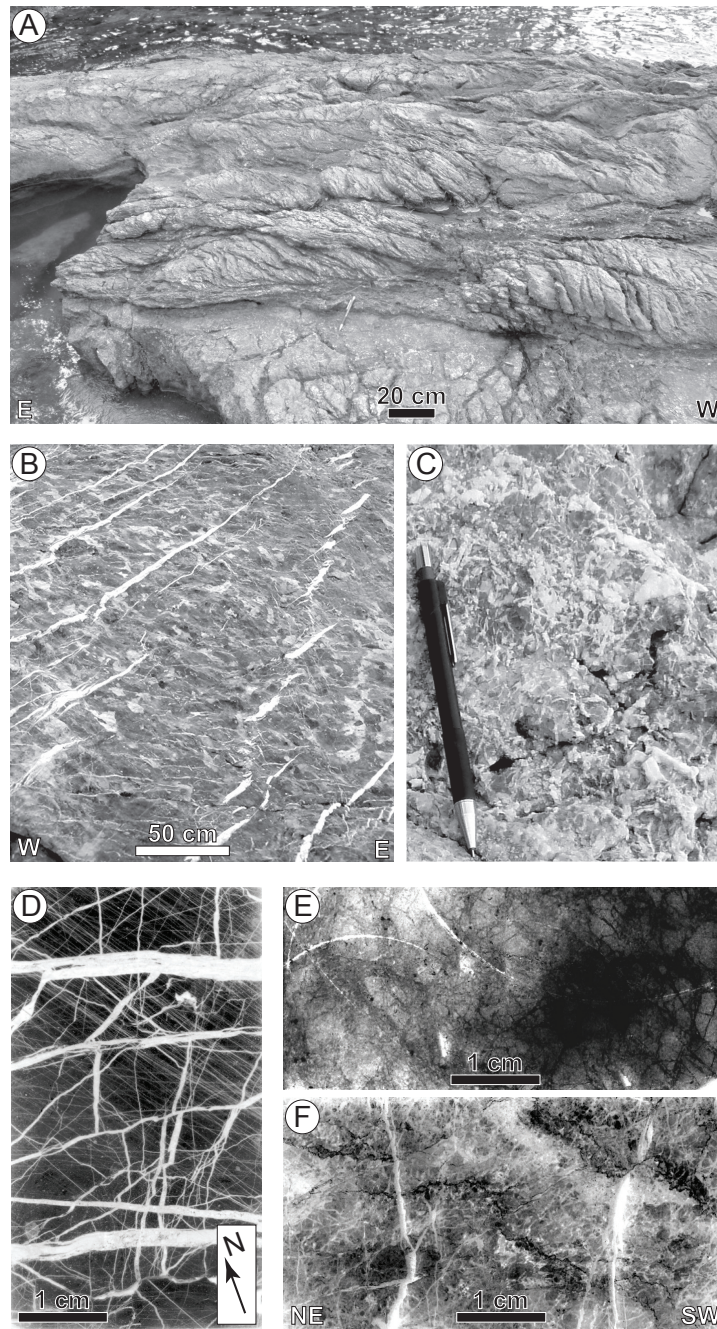


Figure 6.5: Examples of deformation structures associated with the Tellaro Detachment. A) Pressure solution surfaces organized in S-C arrays. B) Calcite veins organized in three systems. B) Randomly oriented calcite filled fractures. C) Differently oriented calcite veins in footwall damage zone in the Biassa Formation (horizontal thin section). E) Dolomitic breccia. F) Dolomitic breccia crosscut by subhorizontal pressure solution surfaces and vertical calcite veins oriented N₃₂₀ (vertical thin section).

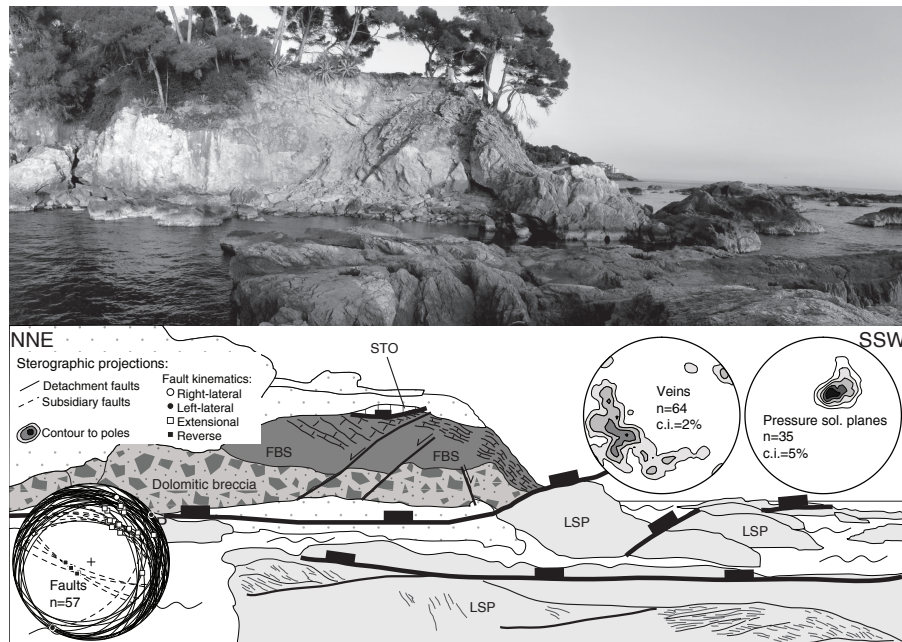


Figure 6.6: Photograph and interpretative line drawing of the Tellaro Detachment at Punta delle Stelle. A dolomitic breccia body is associated with the low-angle shear zone and crosscut by subsidiary high-angle faults. The near horizontal surface in the foreground is affected by intense pressure solution S-C arrays. Structural data collected at this outcrop are provided. See Figure 6.1 for abbreviations used. This outcrop has been affected by a meter-scale rock fall which partly covers the lowermost low-angle slip surface; for a pre-rock fall photo see [Storti \(1995, fig 9c\)](#).

age zones. They have variable orientations (Figures 6.2 and 6.4) and crosscut the carbonate host rocks and the dolomitic breccias. Veins are commonly organized in three sets crosscutting at acute angles (Figure 6.5B): the oldest veins are typically oriented $\sim$ NE-SW, the intermediate $\sim$ E-W and the youngest ones $\sim$ N-S; many exceptions to this general crosscutting chronology are however observed at outcrop scale. At other sites, veins are oriented over a wider range, which covers the strike of the previously described sets and the intermediate directions; crosscutting relationships are even more variable and a systematic relative chronology cannot be established in the field. Finally, adjacent to the low-angle fault zones, calcite filled veins are much shorter, closely spaced, and have irregular and curved geometries (Figure 6.5C).

6.4.2 Petrography and microstructures

Dolomitic breccia bodies are characterized by clast-in-matrix texture. Clasts are dolomitic, with sharp boundaries and very variable crystal size distributions (Figure 6.5E). Subhorizontal pressure solution planes and different generations of small calcite veins crosscut dolomitic breccias (Figure 6.5F). The matrix is fine-crystalline, not resolv-

able with optical microscopy, but dark red / purple CL luminescence indicates a dolomite composition (Figure 6.7A). Calcite veins cut dolomitic breccias with straight boundaries and show no evidence of recrystallization or cement infiltration into the breccia (Figure 6.7A).

Dolomite veins occur only in dolomitic host rocks. They generally have a NW-SE orientation, millimetric to centimetric aperture, and equigranular texture with rhombohedral crystals showing sweeping extinction in crossed-polar microscopy. Crystal size is variable and ranges from $\sim 100\ \mu\text{m}$ to $\sim 1\ \text{mm}$ in thicker veins (Figure 6.7B). Under CL light, dolomite crystals show big cores with dull purple homogeneous luminescence and thin bright purple/red rims at the margins. Occasionally, calcite re-openings are present at the boundary between dolomite veins and dolomitic host.

Petrographic analysis of calcite veins indicates the presence of several successive generations of infillings that can be grouped in two first-order typologies, named “Type 1” and “Type 2”. Type 1 veins show variable aperture (from few tens of μm up to 2 cm), elongated or blocky texture, and are characterized by intensely twinned calcite crystals. Cathodoluminescence petrography highlights the presence of different calcite generations, characterized by none-, dull-brown or bright-orange luminescence and different crosscutting relationships (Figure 6.7C). Multiple re-opening events are also observed in thicker veins (aperture $> 5\ \text{mm}$). Apart from the reopening events, Type 1 veins have homogeneous luminescence colors; among the analyzed samples, only one Type 1 vein is characterized by crystals with growth-zonation luminescence (Figure 6.7D). Type 2 veins are smaller (aperture $< 0.5\ \text{mm}$), have a blocky texture characterized by clean calcite crystals without twinning. Under CL, Type 2 calcite is non-luminescent (Figure 6.7A) or shows a very distinctive non-bright-dull (NBD) growth zonation (Figure 6.7E). Occasionally, small zones of homogeneous dull luminescence are also present in the center of the NBD-zoned veins (Figure 6.7E). Type 2 veins always crosscut or reopen Type 1 vein generations. Systematic investigations, in more than 30 oriented thin sections including many differently oriented veins, indicate that the different generations of calcite infillings do not have a preferred orientation.

6.4.3 *Stable oxygen and carbon isotopes*

Stable isotope analysis was carried out on dolomitic breccia (clasts and matrix), dolomite veins, calcite veins, and host rocks. The cumulative plot of all samples (Figure 6.8A) shows that carbon isotopes are generally clustered while oxygen isotope values show a marked variability ($+1 < \delta^{13}\text{C} < +4\text{‰}$, $-15 < \delta^{18}\text{O} < -2\text{‰}$ V-PDB); only few samples show depleted carbon and less variable oxygen values ($-7 < \delta^{13}\text{C} < +1\text{‰}$, $-7 < \delta^{18}\text{O} < -4\text{‰}$ V-PDB). Comparing different structural positions, samples from the Tellaro Detachment show

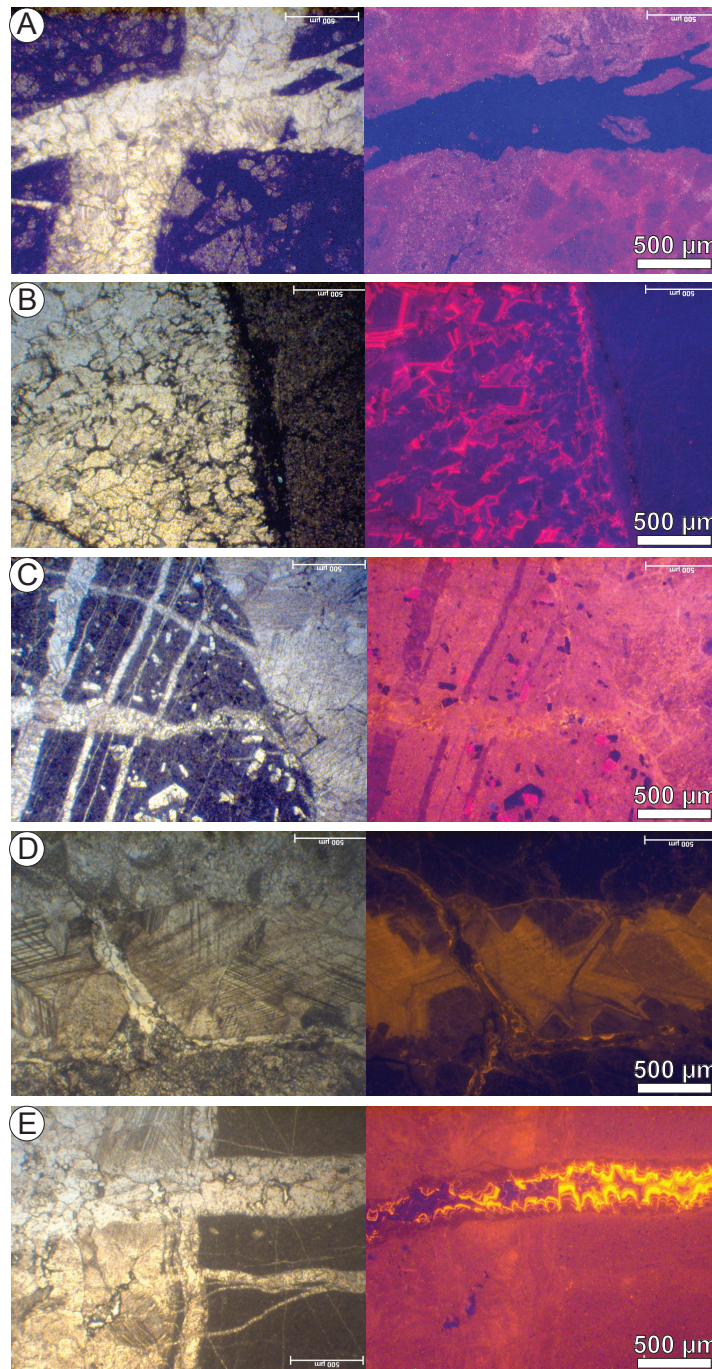


Figure 6.7: Pair of microphotographs (MP) of fault rocks under plane light (left MP) and cathodoluminescence view (right MP). A) Dolomitic breccia crosscut by a “Type 1” calcite vein (vertical in the photo), in turn crosscut by a “Type 2” calcite vein (horizontal in the photo). Note the straight boundaries characterizing the veins, with no evidence of interaction with the surrounding breccia. B) Dolomite vein (left) in dolomitic host rock (right). C) Different generations of “Type 1” calcite veins. D) Type 1 calcite vein with zoned luminescence (sample PB17-1). E) Thick “Type 1” calcite vein (left) characterized by multiple re-openings and crosscut by a thin “Type 2” calcite vein (horizontal), with the typical zoned crystals in cathodoluminescence view.

the greatest oxygen isotope variability, including both the most and the least depleted values. Samples collected far away from major fault zones still show some oxygen isotope variability ($-10 < \delta^{18}\text{O} < -4\text{‰}$ V-PDB), related to the regional background deformation. The only samples showing almost no oxygen isotope variability are those from the Pitelli Detachment ($-7 < \delta^{18}\text{O} < -5\text{‰}$ V-PDB). The latter, however, are by far less numerous.

Along the Tellaro Detachment, host rock values are clustered ($+1 < \delta^{13}\text{C} < +3.5\text{‰}$, $-6 < \delta^{18}\text{O} < -3\text{‰}$ V-PDB), with the exception of one sample with a negative carbon isotope value (Figure 6.8B). A small difference exists between calcite and dolomite hosts, with the latter having slightly more depleted oxygen and enriched carbon isotope values. Dolomitic breccias show clustered carbon and variable oxygen isotopic compositions ($+3 < \delta^{13}\text{C} < +4\text{‰}$, $-7.5 < \delta^{18}\text{O} < -2\text{‰}$ V-PDB). Dolomite veins show slightly variable carbon and clustered oxygen isotopic compositions ($+0.5 < \delta^{13}\text{C} < +3\text{‰}$, $\delta^{18}\text{O} \simeq -3.5\text{‰}$ V-PDB). Calcite vein isotope values are aligned along two trends, one trend with carbon isotope values in the range of the host rock and broadly variable oxygen isotope ($-15 < \delta^{18}\text{O} < -4.5\text{‰}$ V-PDB), the other trend with depleted carbon and less variable oxygen isotope values ($-7 < \delta^{13}\text{C} < +1\text{‰}$, $-7 < \delta^{18}\text{O} < -4\text{‰}$ V-PDB). Such variability in calcite veins shows no correlation with vein orientation or the stratigraphic formations and only a partial correlation with the first order subdivision in successive calcite infill generations (Type 1 or 2; Figure 6.8B). Indeed, veins with depleted carbon ($-7 < \delta^{13}\text{C} < +1\text{‰}$ V-PDB) are preferentially “Type 2”, but veins with more enriched carbon and broadly variable oxygen isotope can be either “Type 1” or “Type 2”.

6.4.4 Fluid inclusion microthermometry

Twenty-five doubly polished wafers of Tellaro Detachment samples were used to analyze fluid inclusions in dolomite, “Type 1” calcite, and “Type 2” calcite veins. Dolomite crystals in dolomite veins can have either cloudy or clean aspect, characterized by more or less abundant fluid inclusions (Figure 6.9, A and B). Inclusions can occur (i) aligned along planes parallel to the crystal margins and different growth sectors, (ii) as planes of pseudosecondary inclusions, or (iii) as isolated spots inside the crystals. Many inclusions are rectangular or prismatic, but shapes that are more irregular are also present. They are variable in size, from less than $1\text{ }\mu\text{m}$ to $5\text{ }\mu\text{m}$, biphasic liquid + vapor at room temperature, with the vapor bubble being 10 to 20% in volume. Microthermometric analysis was carried out on 15 inclusions of primary origin, 2 to $5\text{ }\mu\text{m}$ in size, organized in 4 FIAs, selected from 2 different dolomite veins. Homogenization temperatures (T_h) showed consistent results, clustered at 150°C (Figure 6.10A). First ice melting temperatures (T_{fm}) ranged between -21 and -15°C , but

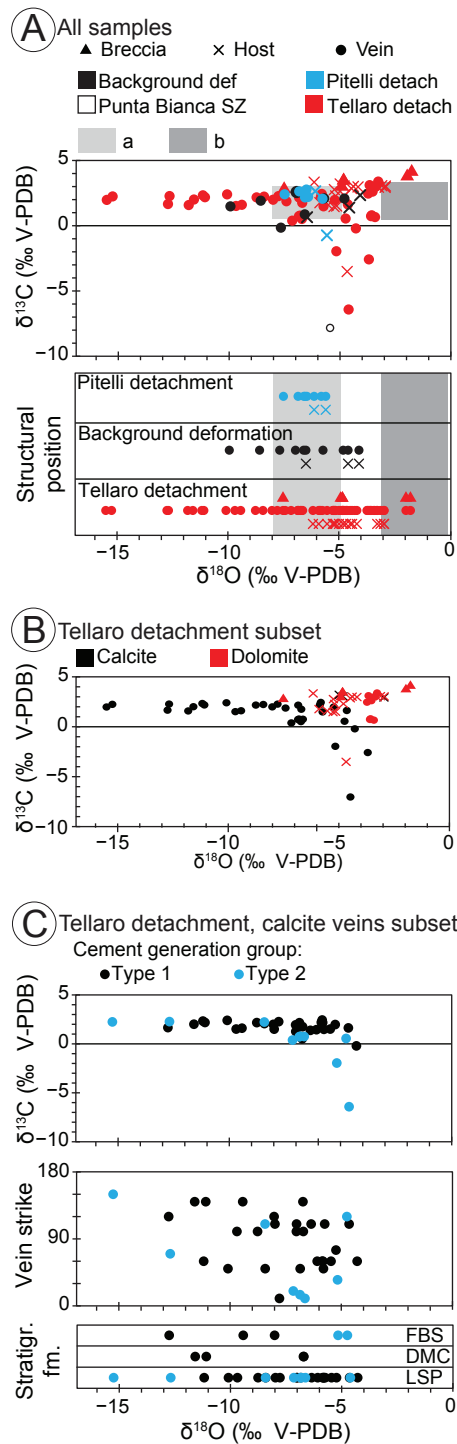


Figure 6.8: Stable isotope data, with different symbols used for host rocks, breccias and veins (original data available in supplementary material). A) Samples from the whole Punta Bianca promontory, with colors indicative of the different structural positions. Reference values of the same host rocks from different areas: a) Cortecchi et al. (1992); b) Gasparrini et al. (2013). B) Tellaro Detachment samples, with colors indicative of sample composition (calcite or dolomite). C) Tellaro Detachment calcite vein samples, with colors indicative of Type 1 and Type 2 generations. Stratigraphic formations: FBS) Biasia Formation; DMC) Dolomie del Monte Castellana Formation; LSP) La Spezia Formation.

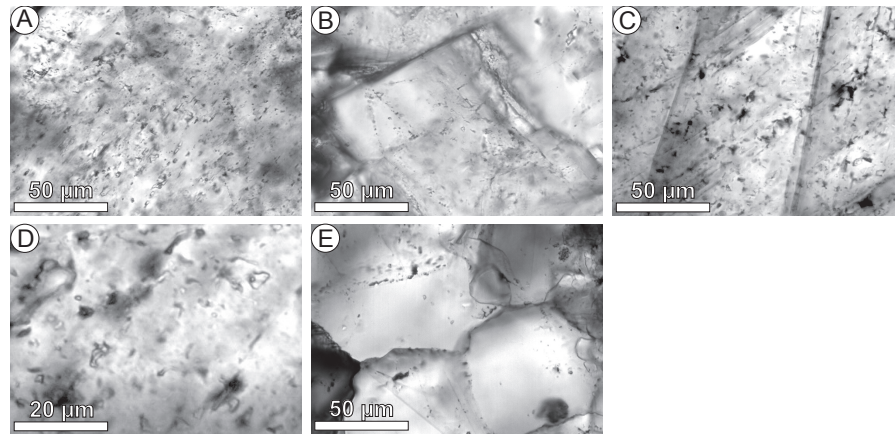


Figure 6.9: Microphotographs of fluid inclusions in carbonate veins. A) Cloudy dolomite crystal with abundant inclusions. B) Clear dolomite crystal with few inclusions. C) Type 1 calcite crystal with secondary inclusions aligned along the twinning planes and isolated inclusions in the space in-between. D) Details of C. E) Type 2 calcite crystal with monophasic fluid inclusions.

they could be measured only in a few inclusions; they were interpreted to indicate a H_2O - NaCl fluid system. Final ice melting temperatures ($T_{\text{m ice}}$) showed a slightly variability (Figure 6.10B), with median value at -15.7°C , corresponding to 19.2% NaCl weight equivalent (Bodnar, 1993).

“Type 1” calcite, due to the intracrystalline deformation and abundant twinning, is mainly characterized by secondary inclusion trails aligned along twinning planes, and by isolated inclusions of primary or unknown origin, randomly distributed in the space between twinning planes (Figure 6.9, C and D). Microthermometric analysis was carried out on the latter, while secondary inclusion trail were not further considered. Isolated inclusions are often too small ($< 1\ \mu\text{m}$) for successful phase change observations, so only samples containing sufficiently large inclusions (2 to $5\ \mu\text{m}$) were selected. Moreover, due to the intense deformation that affected calcite crystals in Type 1 veins, the possible effects of fluid inclusion stretching and leakage had to be evaluated by comparison of data measured in multiple inclusions within the same assemblage. For this reason, veins containing only few (less than 5) isolated inclusions were rejected. Careful petrography and fluid inclusion selection based on above criteria allowed to get reliable microthermometry data from only 4 veins, out of a total of 34 analyzed. The analyzed inclusions are regularly shaped, 2 to $5\ \mu\text{m}$ in size, biphasic liquid + vapor at room temperature, with the vapor bubble being 10 to 20% in volume. Rare inclusions with larger vapor bubbles, up to 60% in volume, were present but not analyzed. Homogenization (T_{h}), first (T_{fm}) and final ($T_{\text{m ice}}$) ice melting temperatures were measured. First ice melting temperatures (T_{fm}) could not be measure in all the inclusions; when available, they ranged be-

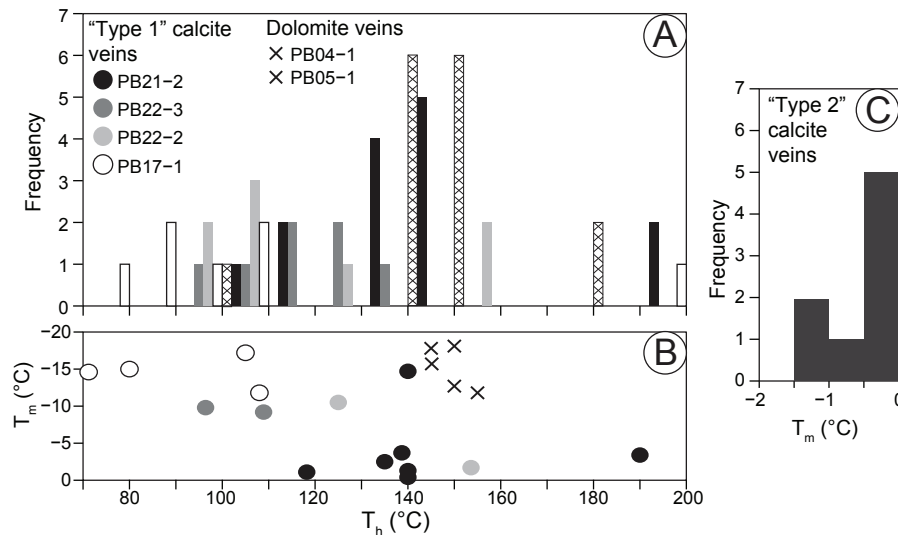
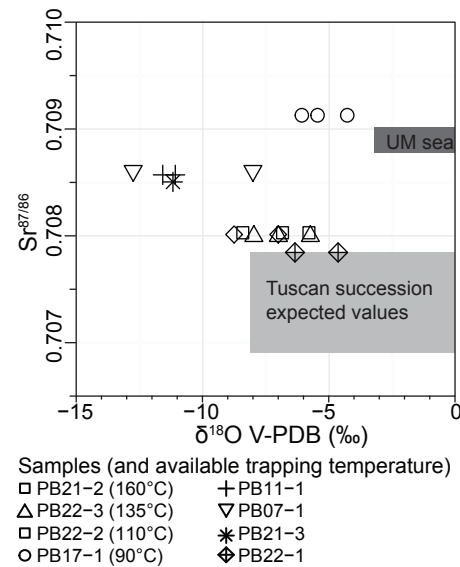


Figure 6.10: Results from fluid inclusion microthermometry (original data available in supplementary material). A) Frequency distribution of homogenization temperatures (T_h) in dolomite (PB04-1, PB05-1) and “Type 1” calcite veins (PB21-2, PB22-3, PB22-2, PB17-1), color-coded according to the veins in which inclusions have been measured. B) Homogenization (T_h) versus final ice melting ($T_{m \text{ ice}}$) temperatures, same samples and color-code as (A). C) Frequency distribution of final ice melting temperatures ($T_{m \text{ ice}}$) in “Type 2” calcite veins.

tween -21 and -15°C , and were interpreted to indicate a H_2O - NaCl fluid system. Homogenization (T_h) and final ice melting ($T_{m \text{ ice}}$) temperatures showed remarkable variability. In vein PB21-2, 14 fluid inclusions were measured in 4 different FIAs. In all assemblages, inclusions showed T_h clustered at 140°C (Figure 6.10A) and relatively clustered $T_{m \text{ ice}}$ with -1.9°C median value (Figure 6.10B), corresponding to 3.2% NaCl weight equivalent (Bodnar, 1993). A single fluid inclusion with a much lower $T_{m \text{ ice}}$ of -14.7°C was however present (Figure 6.10B) and was located within a FIA in which other inclusions showed the more usual $\sim -2^\circ\text{C}$ $T_{m \text{ ice}}$. In vein PB22-2, 8 inclusions were measured in 2 FIAs. In both assemblages, inclusion showed widespread T_h data between 100 and 160°C (Figure 6.10A). Careful analysis, however, indicates that most of the measurements (5) cluster within a small range (92.7 to 107.6°C), while few measurements (3) are remarkably higher and broadly variable (125 to 158.8°C); the latter are located in both FIAs. Reliable $T_{m \text{ ice}}$ data could be measured only in 2 inclusions, located one in each FIA, and gave different values of -10.5 and -1.7°C (Figure 6.10B), corresponding to salinities of 14.4 and 2.9% NaCl weight equivalent (Bodnar, 1993). In vein PB22-3, 7 inclusions were measured in 2 FIAs, showing clustered T_h and $T_{m \text{ ice}}$ values (Figure 6.10A and B), with 120°C and -9.5°C as the most representative and median values, respectively. The latter value corresponds to 13.4% NaCl weight equivalent (Bodnar, 1993). In vein PB17-1, 7 inclusions from 3 FIAs were measured, showing widespread T_h

Figure 6.11: Type 1 calcite veins: radiogenic strontium (one analysis per vein) and stable oxygen (several analyses per vein) isotope compositions (original data available in supplementary material). Grey boxes indicate the expected values for Tuscan succession carbonates and Upper Miocene seawater (see text for details).



data between 80 and 190°C (Figure 6.10A) and clustered $T_{m\ ice}$ data (Figure 6.10B), with median value at -14.8°C, corresponding to 18.5% NaCl weight equivalent (Bodnar, 1993). Careful analysis of the Th values show that most of the measurements (6) are within a relatively small range (80 to 110°C) while only one inclusion have a remarkably higher temperature (190°C).

“Type 2” calcite has a clear aspect, with isolated inclusions of primary or unknown origin. Inclusions are generally small-sized (< 1 µm) but some bigger inclusions (2 to < 5 µm), monophase at room temperature, were found. Such inclusions are of the same size of biphasic inclusions of type 1 calcite and were still monophase after being for 2 weeks at -20°C; they nucleated a bubble only after multiple rapid heating-freezing cycles. This process resulted in biphasic liquid + vapor inclusions, on which standard freezing experiments could be performed. Phase changes associated to the first ice melting could not be observed in any sample. Final ice melting temperature ($T_{m\ ice}$) were measured and showed clustered results with a -0.4°C median value (Figure 6.10C), corresponding to 0.7% NaCl weight equivalent (Bodnar, 1993).

6.4.5 Radiogenic strontium isotopes

Strontium isotope ratios were analyzed from 8 Type 1 calcite veins, which were selected based on the availability of microthermometric data and their stable isotope signature (e.g., those with the most depleted or most enriched oxygen isotope values). For comparison purpose, the $^{87}\text{Sr}/^{86}\text{Sr}$ of analyzed samples is compared with seawater composition at the time of Tellaro Detachment fault activity (late Miocene) and with the expected composition for carbonates of the same age as the Tuscan succession (Upper Triassic to Upper Cretaceous; McArthur et al., 2001). $^{87}\text{Sr}/^{86}\text{Sr}$ values in the analyzed

veins cluster around 3 different values and show a partial correlation with stable isotope signature (Figure 6.11). A first cluster include 4 veins showing depleted oxygen ($-9 < \delta^{18}\text{O} < -4.5\text{‰}$ V-PDB) and $^{87}\text{Sr}/^{86}\text{Sr}$ values around 0.7080, slightly higher than expected values for Upper Triassic to Upper Cretaceous carbonates (e.g., [McArthur et al., 2001](#)); fluid inclusion microthermometry data are available for 3 of those veins and show quite variable homogenization temperatures (110 to 160°C). The second cluster includes 3 veins showing very depleted oxygen isotope ($-12.5 < \delta^{18}\text{O} < -8\text{‰}$ V-PDB) and $^{87}\text{Sr}/^{86}\text{Sr}$ around 0.7085, an intermediate value between late Miocene sea water and expected values for Upper Triassic to Upper Cretaceous carbonates; fluid inclusion microthermometry data are not available for these veins. The last cluster includes only 1 vein showing depleted oxygen isotope values ($-6 < \delta^{18}\text{O} < -4\text{‰}$ V-PDB) and $^{87}\text{Sr}/^{86}\text{Sr}$ values around 0.70915, slightly higher than the values of late Miocene sea water; the homogenization temperatures of the inclusions in this veins are around 90°C.

6.5 DISCUSSION

The activity of the Tellaro Detachment was associated with a complex paleofluid evolution that involved different fluid sources and the precipitation of different generations of mineral infillings over a wide range of temperature and depth conditions. Combined microstructural, petrographic, microthermometric, stable and radiogenic isotope analyses allowed us to identify 3 different main stages within the carbonate precipitation.

6.5.1 *Stages of carbonate precipitation*

The first stage (Figure 6.12A) is characterized by the formation of dolomite veins, in which clustered Th data indicate trapping conditions of $\sim 175^\circ\text{C}$ and 5.2 km (veins PB04-1 and PB05-1). Clustered $T_{\text{m ice}}$ temperatures, indicative of high salinity values, suggest precipitation from a high-salinity fluid. An enriched parent fluid oxygen isotope composition ($\delta^{18}\text{O} \simeq +13\text{‰}$ SMOW; Figure 6.12B) has been calculated from the trapping temperature and the $\delta^{18}\text{O}$ composition of the veins, assuming isotopic equilibrium and using equation 6.1 in the methodology section. It can be proposed that the incipient extensional fault zone was a hydraulically closed system during dolomite vein precipitation, dominated by rock-buffered fluids of local origin, as suggested by (i) the occurrence of dolomite veins only in dolomitic hosts, (ii) the enriched parent fluid signature, and most importantly (iii) the stable isotope composition of the veins which is identical to that of the surrounding rock ([Gray et al., 1991](#); [Mucchez et al., 1995](#)).

The second stage is associated with the formation of type 1 calcite veins. Available microthermometric data show remarkable variability

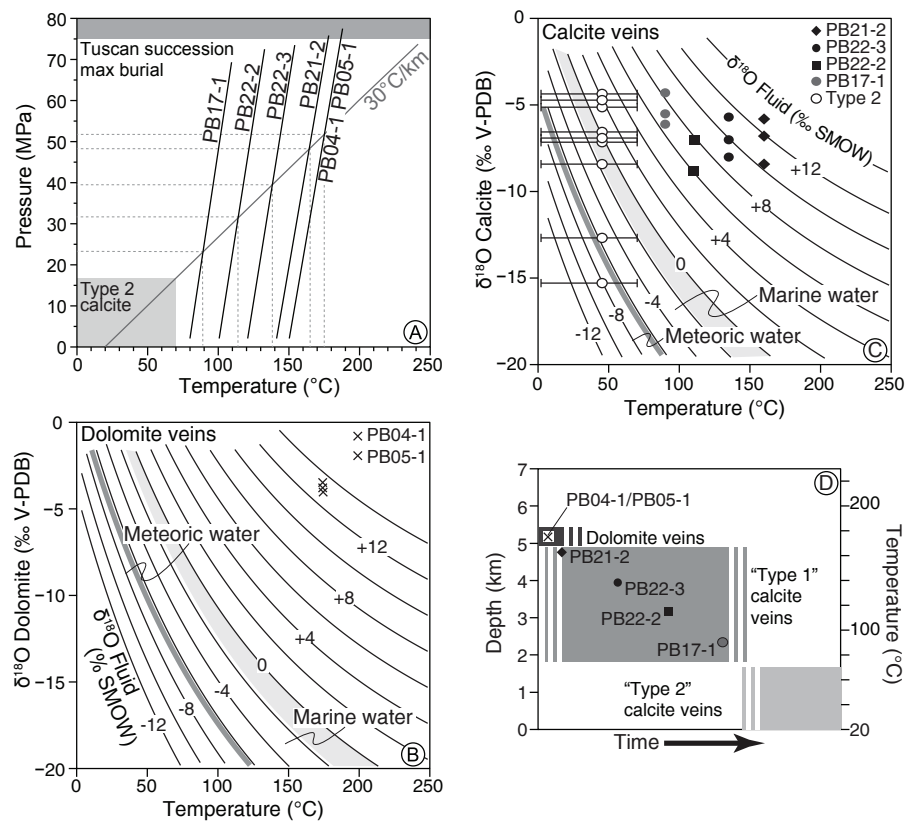


Figure 6.12: A) Pressure correction to estimate the P-T trapping conditions for calcite and dolomite veins. Specific isochores are built, based on to the measured Th and Tm ice data. A geothermal gradient of 30°C/km, in agreement with the gradient that characterized the maximum tectonic burial of the Tuscan succession (Molli et al., 2011), is used as an independent constrain. Impossible conditions are represented by the dark gray area and are constrained by the maximum burial conditions experienced by the Tuscan succession. The two dolomite veins show similar results and are represented by a single isochore; specific isochores are built for each Type 1 calcite vein; Type 2 calcite veins are plotted altogether in the light grey area since their trapping temperature is only constrained as < 70°C. B) Oxygen isotope fractionation during equilibrium precipitation: dolomite (vertical axis) and parent fluid (curves) composition as a function of temperature. Approximate isotopic composition of meteoric and marine water is highlighted in dark and light grey respectively. Black crosses indicate dolomite veins. C) Oxygen isotope fractionation during equilibrium precipitation: calcite (vertical axis) and parent fluid (curves) composition as a function of temperature. Black and gray symbols indicate the different Type 1 calcite veins; white circles indicate Type 2 calcite veins. Note that the error bar in Type 2 calcite veins is indicative of the weak constrain available for the trapping temperature ($T < 70^\circ\text{C}$, see text for details). D) Schematic diagram summarizing the timing and depth conditions of the different carbonate precipitation stages (see text for details). Symbols indicate Dolomite and type 1 calcite veins as in B and C.

in this stage. However, individual data from each vein cluster within reasonable T_h ranges, with the exception of few inclusions affected by leaking or stretching (see below). Therefore, the variable T_h are interpreted to reflect the precipitation of each vein infilling at a different temperature. Such observation can imply that (i) each vein formed at a different depth, and precipitation temperature was controlled by a regional geothermal gradient, or alternatively (ii) all veins formed nearly at the same depth, and calcite precipitation temperature was controlled by the parent fluid temperature (e.g., in a hydrothermal or frigidothermal system). The first hypothesis will be used for data interpretation, because is favored by several evidence, including: i) there is not a systematic relationship between homogenization temperatures and fluid salinity, despite the presence of both high-salinity and low-salinity fluid in the different veins; ii) within a single vein, inclusion with very different salinity homogenize at the same temperature. Using these assumptions, and available microthermometric data, different temperature-depth conditions have been estimated for type 1 calcite veins. Vein PB21-1 is the calcite vein characterized by the deepest and hottest trapping conditions: clustered homogenization temperatures (Figure 6.10A) are compatible with vein formation at $\sim 165^\circ\text{C}$ and 4.8 km depth (Figure 6.12A). Relatively clustered $T_{m\text{ ice}}$ values indicate the main involvement of a low-salinity fluid, but the presence of an inclusion with a much lower value provides evidence for a possible mixing with a high-salinity fluid. In vein PB322-3, clustered microthermometric data (Figure 6.10A) are compatible with vein formation at $\sim 140^\circ\text{C}$ and 4 km depth (Figure 6.12A), from a high-salinity fluid. In vein PB22-2, the three inclusions with higher T_h s are considered to be affected by leakage or stretching and not related anymore with the original trapping conditions. Consequently, the clustered T_h s of the remaining inclusions (Figure 6.10A) are used to estimate $\sim 115^\circ\text{C}$ and 3.2 km trapping conditions (Figure 6.12A). Fluid mixing is suggested by the variable $T_{m\text{ ice}}$ s (hence salinity values) corresponding to assemblages trapped at the same temperature, within the same vein (Figure 6.10B). Similarly, in vein PB17-1 the inclusion with the highest T_h is considered to be affected by leakage or stretching and not related anymore with the original trapping conditions. However, in this vein, the remaining inclusions might have been affected by minor stretching or leaking, which led to T_h variability over a relatively wide range (75 to 110°C ; Figure 6.10A). To minimize this effect, the trapping conditions of this vein are estimated from the lowest T_h range; applying the usual pressure correction to such value allows estimating trapping conditions of $\sim 90^\circ\text{C}$ and 2.3 km (Figure 6.12A). Within this vein, $T_{m\text{ ice}}$ s are clustered in all the inclusions (Figure 6.10B) and indicate a high-salinity fluid. The parent fluid isotopic composition has been calculated from the available trapping temperatures and the $\delta^{18}\text{O}$ composition of the veins, assuming isotopic equilibrium and using equation 1 in the methodology sec-

tion. It resulted an enriched oxygen isotope composition ($+6 < \delta^{18}\text{O} < +12\text{‰}$ SMOW) for all “Type 1” calcite veins with well-constrained trapping conditions (Figure 6.12C).

Beside salinity values variability, the presence of different fluids involved in “Type 1” calcite veining is confirmed by petrographic observations, indicating the presence of multiple generations of calcite infill within the different type 1 veins (Figure 6.7C and D), and by the different $^{87}\text{Sr}/^{86}\text{Sr}$ (Figure 6.11). Moreover, Sr isotope composition shows a partial correlation with the stable oxygen isotope values, indicating that veins with the most depleted oxygen ($-12.5 < \delta^{18}\text{O} < -8\text{‰}$ V-PDB) might have formed from a specific fluid. Other notable characteristics of type 1 calcite veins are: i) the absence of systematic cross-cutting relationships, indicating the absence of a relative chronology between differently oriented veins; ii) the absence of a preferred orientation for each calcite generation. Both these observations can be easily explained if differently oriented veins opened at about the same time. This process has been already described in similar settings, and is favored by low-differential stress and high-fluid pressure conditions (Collettini et al., 2006a).

The third stage of carbonate precipitation is associated with “Type 2” calcite veining. The absence of twinning in type 2 calcite crystals indicates that they have not been affected by tectonic strain and supports their formation in a post-kinematic frame, when the Tellaro Detachment fault system was no longer active and almost completely exhumed. Fluid inclusions in “Type 2” calcite are monophasic at room temperature and do not allow to determine the trapping temperature via traditional heating experiments. However, it can be proposed that such monophasic inclusions are indicative of low-temperature precipitation ($T < 70^\circ\text{C}$) because: i) they are of the same size as the biphasic inclusions in other calcite veins; ii) they did not nucleate a vapor bubble after being at -20°C for 2 weeks (Goldstein & Reynolds, 1994). Such trapping temperature would correspond to < 1.7 km depth conditions (Figure 6.12A). The low salinity values corresponding to the high $T_{\text{m ice}}$ measured, and the zoned NBD luminescence that is common in “Type 2” calcite are compatible with a parent fluid of meteoric origin. Based on the stable isotope geochemistry, “Type 2” calcite veins can be subdivided into two groups (Figure 6.8C): i) veins with depleted carbon and moderately depleted oxygen ($-7 < \delta^{13}\text{C} < +1\text{‰}$, $-7 < \delta^{18}\text{O} < -4\text{‰}$ V-PDB); ii) veins with enriched carbon and very depleted oxygen ($+1 < \delta^{13}\text{C} < +3\text{‰}$, $-7 < \delta^{18}\text{O} < -15\text{‰}$ V-PDB). In the former, low-temperature precipitation in a shallow environment is suggested by: i) the depleted carbon, indicative of precipitation from meteoric water influenced by soil-derived CO_2 ; ii) the parent fluid oxygen signature, which is similar to the meteoric water signature in the area (Allan & Matthews, 1977; Videtich & Matthews, 1980). In the latter group, the very depleted oxygen might be compatible with precipitation at slightly higher temperatures, while the

carbon is rock-buffered, indicating fluid-rock interactions (Walkden & Williams, 1991; Muchez et al., 1991).

In addition to the above stages, a dolomitization event also affected the fault zones and led to development of the tabular-shaped bodies associated to the major low-angle fault segment (Figures 6.4 and 6.6). Structural observations locate this dolomitization event in the early history of fault activity; indeed, with progressive deformation, dolomitic bodies are brecciated and systematically crosscut by high-angle subsidiary faults, subhorizontal pressure-solution planes and calcite veins. However, the exact timing and temperature-depth conditions of the dolomitization event are not constrained. Moreover, it is important to note that direct crosscutting relationships between dolomite vein (fluid stage 1) and type 1 calcite veining (fluid stage 2) are sparse and do not allow full constraint of the relative timing. Possibly, both stages started nearly at the same time and coexisted for a while during early fault activity. Afterwards, with continuing deformation, dolomite veining stopped, and type 1 calcite veining continued for a long period allowing vein precipitation over variable temperature-depth conditions (Figure 6.12D).

6.5.2 *Structural and paleofluid evolution*

The kinematic consistency of fault slickenlines, S-C fabric, fault corrugations, synthetic and antithetic secondary fault zones, combined with multiple cross-cutting relationships of vein systems and the variable trapping temperature estimated for the different fault-related veins, collectively indicate that the Tellaro Detachment was active as a low-angle extensional fault system over a long-time period, during progressive exhumation of the Tuscan basinal succession (Xu, 1997; Crispini & Frezzotti, 1998; Boullier, 1999; Kenis et al., 2000; Essarraj et al., 2001; Montomoli et al., 2001; Fischer et al., 2009; Jacques et al., 2014). Combining field and laboratory data, we propose that the structural and paleofluid evolution of the fault system can be schematically divided in four main deformation events: i) early fracturing and incipient faulting, at approximately 5 km depth, characterized by circulation of rock-buffered fluids of local origin in a closed system environment (Figure 6.13A); ii) dolomitization of the fault zone, presumably at similar depths, and resulting self-sealing of the major fault zones that favored footwall fluid overpressures (Figure 6.13B); iii) main activity of the fault system, including brecciation, cementation and faulting of the dolomitized bodies; at this stage the large-scale fault system extended through the upper 5 km of the brittle crust, thus promoting the development of highly permeable damage zone that provided an effective pathway for fluid migration and mixing in an open hydraulic system (Figure 6.13C); iv) exhumation to shallow depth, deactivation and possible passive tilting of the Tellaro Detachment fault system by footwall isotactic adjustment (e.g., Buck,

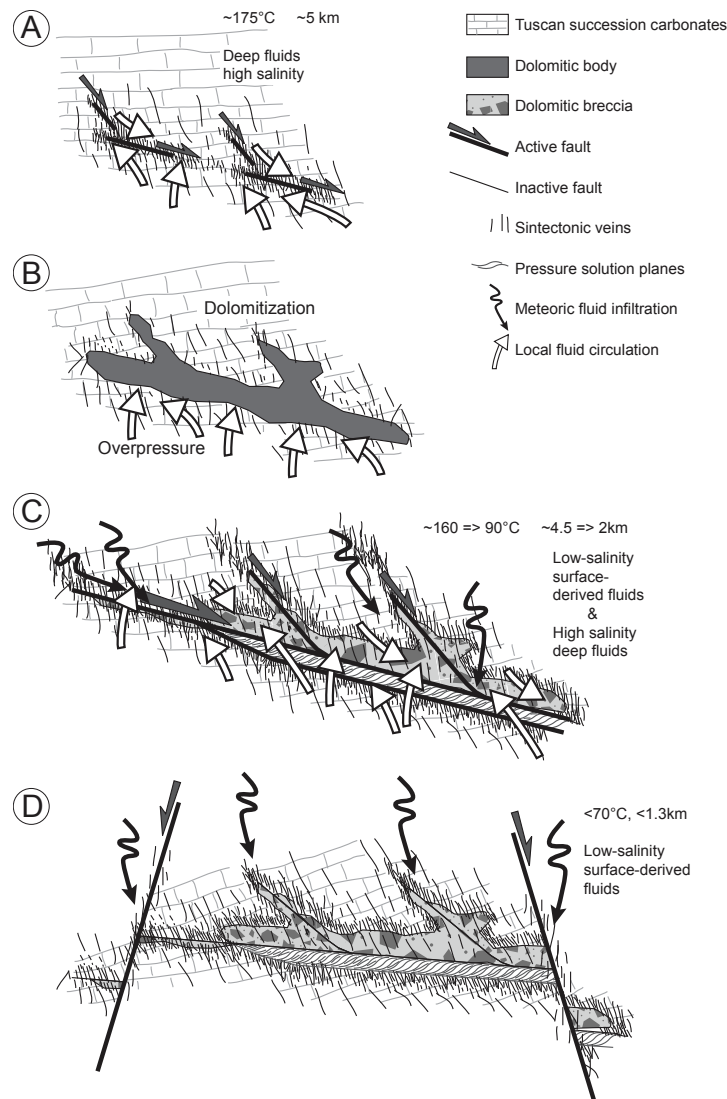


Figure 6.13: Schematic cartoon illustrating the proposed structural and paleo-fluid evolution of the Tellaro Detachment into four deformation events (see text for details).

1988) and superimposed high-angle faulting. Residual fault zone permeability at this stage allowed the circulation of meteoric fluids (Type 2 calcite veining, Figure 6.13D). The timing of the Tellaro Detachment activity can be inferred from regional geological constraints. The activity of the high-angle faulting (Figure 6.13D) started in Pliocene, as constrained by the stratigraphy of the associated basin infillings (Federici, 1973; Federici & Rau, 1980; Bernini et al., 1990; Moretti, 1992; Martini et al., 2011). The maximum tectonic burial of the Tuscan succession in the area of interest is dated to the middle Miocene (Fellin et al., 2007, and references therein). The main activity of the Tellaro Detachment (Figure 6.13A to C) is therefore late Miocene, because it postdates the maximum tectonic burial of the Tuscan Succession and is overprinted by the high-angle faulting.

The proposed model involves only possible gentle tilting of late Miocene low- and high-angle faults during Plio-Pleistocene deformations. Despite a lack of appropriate evidence for a paleohorizontal reference surface, such an inference can be made based on the attitude of high-angle subsidiary faults in the damage zone. The maximum amount of possible tilting can be constrained by the assumption that subsidiary extensional faults formed as conjugate pairs bisected by the vertical principal axis of the stress ellipsoid σ_1 . Since the present-day dip of synthetic and antithetic extensional faults is $\sim 35^\circ$ and 70° toward the NE and SW respectively (Figure 6.2), a southwestward rotation of 17° is necessary to restore them to a symmetrical configuration characterized by average slip surfaces dipping $\sim 52^\circ$ toward both the NE and the SW. Such titling implies that the late Miocene master slip surfaces of the Tellaro Detachment had an original average dip of 26° toward NE, a value that fits well into the range of low-angle normal faults described in extensional continental settings worldwide, including the modern cases of the Corinth Gulf (e.g., [Rietbrock et al., 1996](#)), Papua New Guinea (e.g., [Abers et al., 1997](#)) and the Alto Tiberina Fault (e.g., [Chiaraluce et al., 2007](#)). However, the evidence that subsidiary faults developed in the footwall damage zone of the master detachment questions the assumption of a vertical σ_1 . Indeed, in case of simple shear, a kinematically-induced stress field (σ_{1k}) is expected to be generated during slip, in the proximity of master faults. Such σ_{1k} is oriented at 45° synthetically from the principal slip surface (e.g., [Harding, 1976](#); [Logan et al., 1973](#); [Storti et al., 2006](#)). The vectorial summation of the kinematically induced σ_{1k} and the gravity related vertical σ_1 produces a resultant stress field that has the principal axis inclined in the tectonic transport direction by an angle greater than 45° .

In the case of the Tellaro Detachment, this implies that (i) the dip of the southwestward-dipping antithetic faults was greater than that of the northeastward dipping ones, and (ii) the average dip of the low-angle principal slip zone was lower than 26° . The most likely mechanical reason supporting slip on such a misoriented extensional detachment is fluid overpressuring ([Miller et al., 1996](#); [Collettini & Sibson, 2001](#)) as also supported by widespread evidence of hydraulic fracturing preserved in vein arrays adjacent to the principal slip zones (Figure 6.5C). At a large-scale, this low-angle fault zone affecting the Tuscan succession can be interpreted as the flat segment of a staircase crustal-scale extensional fault system affecting the shallower portion of the Northern Apennines orogenic wedge.

6.6 CONCLUSIONS

The Tellaro Detachment provides a well-exposed example of a low-angle extensional fault system affecting carbonate rocks, characterized by long-lasting kinematic activity during progressive exhumation.

tion from ~ 5 km depth. The original average dip of the principal slip zone of the Tellaro Detachment is estimated to be lower than 26° toward the northeast, based on conjugate fault pairs in damage zones. As a consequence, our data support low-angle normal faults as pristine deformation structures, not affected by significant post-kinematic passive rotation.

Fluid circulation was associated with fault activity and resulted in the precipitation of different generations of dolomite and calcite infillings in fault breccias and syntectonic veins, at gradually decreasing temperature and depth conditions. Initial incipient faulting was characterized by circulation of rock-buffered local fluids in a "closed system" environment. Subsequently, the propagation of the fault system, and the associated damage zone permeability, favored mixing of different fluids in an "open system" environment. Elevated fluid pressures and low differential stress conditions favored shear localization along the misoriented low-angle principal slip zones and contributed to the precipitation of calcite infillings in differently oriented veins in damage zones, as well as hydrofracturing processes in the immediate proximity of the principal slip zones themselves.

Large-scale low-angle normal faults, similar to that described in the paper, are common in many orogens worldwide and can be associated with processes of notable significance such as seismic hazard, formation of ore deposits and hydrocarbon trapping. Structural, paleohydrological, geochemical and microthermometric information from Tellaro Detachment can provide useful insights for better characterization of comparable fault systems in the subsurface.

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SUPPLEMENTARY MATERIAL

Table 6.1: Oxygen and carbon stable isotopes composition

Sample	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
PBo1-1	-3.598015627	0.733227041
PBo1-2	-3.419678856	0.663677359
PBo1-Host	-4.18635566	2.973466285
PBo3-1	-3.729762819	2.463321582
PBo3-2	-3.379600337	3.00152199
PBo3-3	-7.417394571	1.874805795
PBo3-Host	-3.118391304	3.065388078
PBo4-1	-3.530494971	2.614875097
PBo4-2	-3.57111651	0.797864052
PBo4-Host	-6.166782156	3.339248957
PBo5-1	-3.663816209	3.109570468
PBo5-Host	-4.399261101	3.012282152
PBo7-1	-8.018585139	1.99257591
PBo7-1	-12.75920964	1.662641494
PBo7-2	-5.165024312	-1.955821058
PBo7-3	-3.697015858	-2.579376966
PBo7-Host	-5.150854041	1.510967941
PBo8-1	-4.746681892	0.552615263
PBo8-2	-9.428697734	1.603749477
PBo8-3	-11.82397817	1.584601039
PBo8-Host	-5.26845378	1.468712742
PBo9-Breccia (clast)	-4.916475977	3.033797594
PBo9-Breccia (clast)	-1.968848867	3.822461348
PBo9-Breccia (matrix)	-1.766230243	4.17329713
PBo9-Breccia (matrix)	-4.836281518	3.510388468
PBo10-1	-6.721876568	0.518048535
PBo10-Breccia (clast)	-7.519001058	2.848463837
PBo10-Breccia (matrix)	-4.836281518	3.510388468
PBo11-1	-6.7048689	1.755540169
PBo11-2	-11.58865457	2.00860064
PBo11-3	-11.0915862	2.183966649
PBo11-Host	-5.905422272	1.767836275
PBo12-1	-5.447715286	-7.830986049

Table 6.1: continues in next page

Table 6.1: continues from previous page

Sample	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
PB13-1	-3.275979002	3.383419409
PB13-2	-5.861382483	2.202815157
PB13-3	-7.778508883	2.267879378
PB13-Host	-2.965683941	2.935542809
PB14-1	-7.155310166	0.386850382
PB14-2	-6.631690314	0.753067916
PB14-Host	-2.952480673	3.03880113
PB14-Host	-3.266847105	2.950860403
PB15-1	-10.09282651	2.403239101
PB15-2	-15.25664955	2.245406065
PB15-Host	-5.253281872	2.771902766
PB16-1	-5.236265379	1.975000714
PB16-2	-6.852577244	0.728559568
PB16-3	-6.806799774	0.772631282
PB16-Host	-4.48261693	2.852195349
PB16-Host	-4.676277174	-3.509995109
PB17-1	-4.283019794	-0.20557617
PB17-1	-5.459353413	1.515126207
PB17-1	-6.08061525	1.449258294
PB17-Host	-4.990328454	3.199481177
PB18-1	-6.576471151	2.171014854
PB18-1	-6.44002788	2.209491878
PB18-Host	-5.589577663	-0.735877433
PB19-1	-6.515702392	2.771270875
PB19-2	-5.792735767	2.065401557
PB19-3	-6.838990884	2.631945064
PB19-4	-7.49867515	2.409456903
PB19-Host	-6.114486804	2.643351011
PB20-1	-12.6889857	2.27345181
PB20-2	-9.689896758	1.514739749
PB20-3	-8.424521977	2.230643413
PB20-Host	-4.797043395	2.304042128
PB21-1	-5.822191176	2.437755202
PB21-2	-6.845528372	2.163001815
PB21-2	-5.799519954	2.177478739

Table 6.1: continues in next page

Table 6.1: continues from previous page

Sample	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
PB21-2	-8.419625285	2.073103431
PB21-3	-11.18502527	2.33031912
PB21-Host	-5.196944731	2.287167823
PB22-1	-4.644876452	1.631109287
PB22-1	-6.352577593	1.381870352
PB22-2	-8.762716221	2.172404502
PB22-2	-7.016462199	1.940908956
PB22-3	-5.743221488	1.486152691
PB22-3	-7.008640506	1.265714821
PB22-3	-7.980512419	1.504866013
PB22-Host	-5.528908758	1.593137589
PB23-1	-5.712871801	2.098031229
PB23-2	-7.667225462	-0.148362865
PB23-3	-8.575878709	1.917846138
PB23-Host	-4.581880439	1.388985571
PB24-1	-9.942800053	1.4847089
PB24-2	-4.8018261	2.083882079
PB24-3	-6.955664199	2.583000556
PB24-Host	-4.089951326	2.33853068
PB25-1	-6.588829163	0.871188051
PB25-Host	-6.484915477	0.652255224
PB26-1	-4.612401377	-6.418925441
PB26-2	-15.51976416	1.979443476
PB26-Host	-4.816208721	3.042198218

Table 6.1: ends from previous page

Table 6.2: Strontium isotope contents

Sample	$^{87}\text{Sr}/^{86}\text{Sr}$	% Std Err
PB07-1	0.708623	0.0012
PB11-2	0.708571	0.0013
PB17-1	0.709129	0.0012
PB21-2	0.708029	0.0011
PB21-3	0.708505	0.0015
PB22-1	0.707845	0.0017
PB22-2	0.708013	0.0013
PB22-3	0.707983	0.0012

Table 6.3: Fluid inclusion microthermometry

Sample		FIA	T _h	T _{freezing}	T _{fm}	T _{m ice}
PB04-1	Dolomite vein	A	150	-73.2	-25.2	-12.7
PB04-1	Dolomite vein	A	180	-46.4	-20.7	
PB04-1	Dolomite vein	A	155			
PB04-1	Dolomite vein	A	155			
PB04-1	Dolomite vein	C	145			-17.8
PB04-1	Dolomite vein	C	182			
PB05-1	Dolomite vein	A	145			-15.7
PB05-1	Dolomite vein	A	140			
PB05-1	Dolomite vein	A	100.4			
PB05-1	Dolomite vein	A	155			-11.8
PB05-1	Dolomite vein	A	155			
PB05-1	Dolomite vein	A	145			
PB05-1	Dolomite vein	C	140			
PB05-1	Dolomite vein	C	150	-71.8		-18.1
PB05-1	Dolomite vein	C	140			
PB21-2	Type 1 calcite vein	A	140			
PB21-2	Type 1 calcite vein	B	135	-38.4		-2.5
PB21-2	Type 1 calcite vein	B	140	-37.5	-20.8	-0.4
PB21-2	Type 1 calcite vein	B	140	-37.5	-19.2	-1.3
PB21-2	Type 1 calcite vein	B	135			
PB21-2	Type 1 calcite vein	B	106.6			
PB21-2	Type 1 calcite vein	B	140			

Table 6.3: continues in next page

Table 6.3: continues from previous page

Sample		FIA	T _h	T _{freezing}	T _{fm}	T _{m ice}
PB21-2	Type 1 calcite vein	B	135			
PB21-2	Type 1 calcite vein	C	140	-60	-24.3	-14.7
PB21-2	Type 1 calcite vein	C	138.7	-34.7	-19.1	-3.7
PB21-2	Type 1 calcite vein	G	118.2	-41.6	-22.1	-1.1
PB21-2	Type 1 calcite vein	G	190			
PB21-2	Type 1 calcite vein	G	115.7			
PB21-2	Type 1 calcite vein	G	190			-3.4
PB22-2	Type 1 calcite vein	A	95.7			
PB22-2	Type 1 calcite vein	A	153.6	-44.6	-18.8	-1.7
PB22-2	Type 1 calcite vein	A	103.8			
PB22-2	Type 1 calcite vein	A	103.5			
PB22-2	Type 1 calcite vein	B	92.7			
PB22-2	Type 1 calcite vein	B	107.6			
PB22-2	Type 1 calcite vein	B	125	-68.2	-19.1	-10.5
PB22-2	Type 1 calcite vein	B	158.8			-21.8
PB22-3	Type 1 calcite vein	D	96.4		-13.3	-9.8
PB22-3	Type 1 calcite vein	D	108.9			-9.2
PB22-3	Type 1 calcite vein	E	120			
PB22-3	Type 1 calcite vein	E	115			
PB22-3	Type 1 calcite vein	E	115			
PB22-3	Type 1 calcite vein	E	120			
PB22-3	Type 1 calcite vein	E	130			
PB17-1	Type 1 calcite vein	A	71.2	-64	-20.8	-14.6
PB17-1	Type 1 calcite vein	A	80			
PB17-1	Type 1 calcite vein	A	80	-76	-20.4	-15
PB17-1	Type 1 calcite vein	C	108		-19.7	-11.8
PB17-1	Type 1 calcite vein	C	190		-24.2	
PB17-1	Type 1 calcite vein	D	95			
PB17-1	Type 1 calcite vein	D	105	-73		-17.2
PBo7-3	Type 2 calcite vein					-1.4
PBo7-3	Type 2 calcite vein					-0.2
PBo7-3	Type 2 calcite vein					-0.4
PBo8-1	Type 2 calcite vein					-0.1
PBo8-1	Type 2 calcite vein					-1.3
PB15-3	Type 2 calcite vein					-0.4

Table 6.3: continues in next page

Table 6.3: continues from previous page

Sample	FIA	T _h	T _{freezing}	T _{fm}	T _{m ice}
PB15-3 Type 2 calcite vein					-0.3
PB16-3 Type 2 calcite vein					-0.7

Table 6.3: ends from previous page

LUNIGIANA FAULT SYSTEM

The regional-scale deep structural architecture of the Lunigiana and Tuscan-Emilian sectors of the Northern Apennines have been studied through the interpretation of commercial 2D seismic profiles. Structural interpretation of available subsurface data, integration with field geology, and information from the literature allowed us to build and validate a 3D structural model of this sector of the Northern Apennines.

A confidentiality agreement prevents the possibility to illustrate the geophysical dataset used in this study and, consequently, constitutes a severe limitation to the possibility of publishing the results of our work. This is the main reason for the limited length of this chapter. Nevertheless, our results provide a first-order characterization of the large-scale structures of this sector of the belt which is useful to locate the Val di Lima fault system and Tellaro Detachment case studies (chapters 5 and 6) in their regional framework.

7.1 SUBSURFACE STRUCTURES

Investigated subsurface data include few wells and many 2D seismic profiles, NW-SE and SW-NE oriented, which provide a good coverage of the area of interest (Figure 7.1).

The interpreted profiles show remarkable similarities and analogies all over the study area (Figure 7.2). Geological interpretation of subsurface data indicates the presence of different structures in the internal and central sectors of the belt; therefore, those two sectors will be described separately.

7.1.1 *Internal sector*

The internal portion of the Apenninic belt is characterized by a well-developed seismic reflector (r) which is sub-horizontal and located at shallow depths (3 to 5 km). Such a reflector has been drilled by the Pontremoli 1 well (p1) which revealed the presence of metamorphic basement rocks below the reflector (r) itself (Progetto ViDEPI, 1971). The upper portion of the thrust wedge, above reflector (r), is characterized by extensional fault systems associated with the present-day morphology of the Lunigiana and Vara valleys (al and vv, respectively), limited by NE and SW dipping high-angle fault zones (Bernini, 1991). The lower portion of the wedge, below the reflector (r) is characterized by anastomosed structures that can be interpreted as compressional structures that delimitate NE-verging thrust

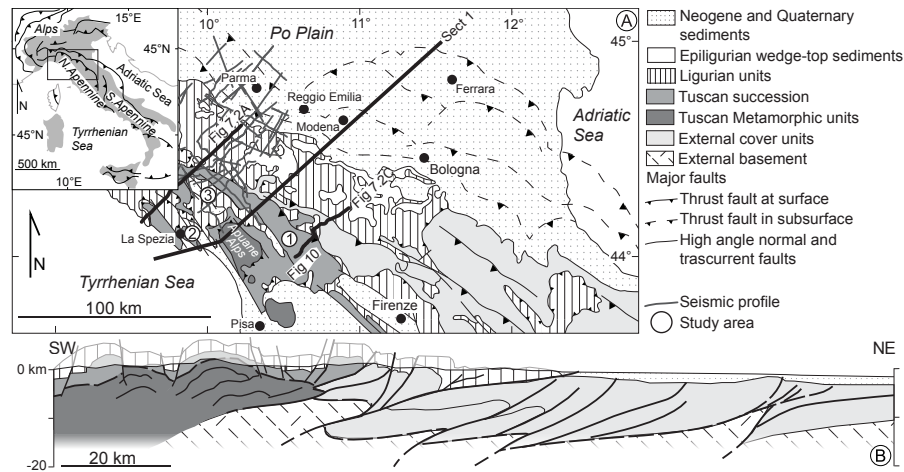


Figure 7.1: Regional setting of the area of interest. A) Tectonic map of the Northern Apennines, (modified from [Molli, 2008](#)) and location of the case studies investigated in this Thesis: 1) Val di Lima fault system; 2) Tellaro Detachment; 3) Lunigiana fault system. B) Regional-scale geologic cross section (modified from [Molli, 2008](#)).

sheets, sometimes characterized by lateral-ramp terminations toward the NW. In the innermost sector of the belt, in the proximity of the Tyrrhenian margin, few SW-vergent backthrust are present as well (b1-b3 in Figure 7.2A and B). The backthrusts cut the (r) reflector and compressively determine a ~4 km offset.

7.1.2 Central sector

In the central sector of the belt, the prosecution toward the NE of the regional seismic reflector (r) is characterized by a different attitude. Indeed, it is moderate dipping and gradually deepens toward the NE till disappearing at the bottom limit of seismic dataset. Above the reflector (r), the belt is characterized by the presence of several NE-vergent blind thrusts (t1 - t5) cutting at high angle the tectonic contact between major tectonic units (Ligurian and Subligurian units, Tuscan succession and External cover units). Blind thrusts have average dip (~ 25 to 40°), variable offset (~ 500 to 1500 m) and their tips are located at shallow depths within the Ligurian and Subligurian units. The along-strike length of the thrust sheets is sometimes difficult to estimate because of the poor quality of the available seismic data. The most external thrust sheet (t5) seems to be the larger one, being present through the whole study area. The most internal thrust sheet (t1), associated with the front of the Tuscan succession, is the smaller one; indeed, it is characterized by a lateral-ramp termination toward the NW in correspondence of the Pracchiola tectonic window. Moreover, in the proximity of the Apennines topographic divide, at the border between the internal and central sector of the belt, the SW-vergent high-angle extensional faults limiting the Lunigiana valley

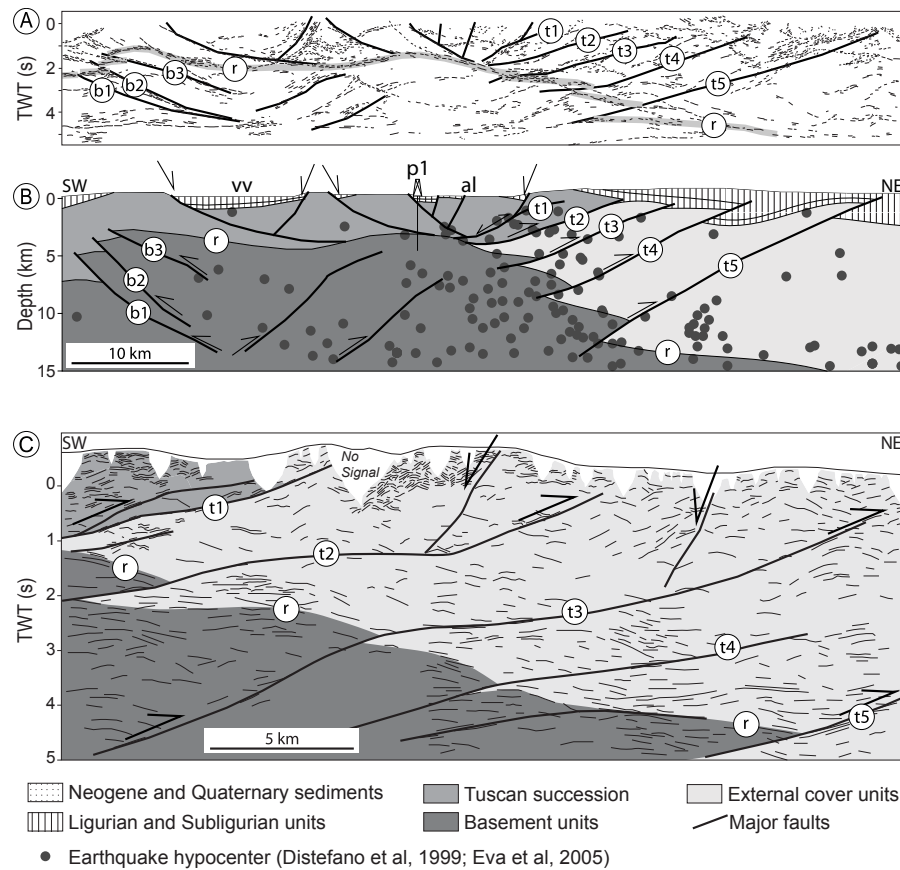


Figure 7.2: Commercial seismic profiles through the Northern Apennines (see Figure 7.1 for location and text for details). A) Line-drawing of the Levanto-Parma seismic profile (after Camurri, 2000) and structural interpretation. B) Depth-conversion of (A) and current seismicity (after Di Stefano et al., 1999; Eva et al., 2005). C) Line-drawing and geological interpretation of a seismic profile locate in the proximity of the Lima valley.

sole down on the (t₁) thrust fault and determine a negative inversion of its deeper portion.

At depth, the rooting of the thrust faults and their relationships with the reflector (r) are variable. Sometimes the faults cut the reflector (r) determining a notable offset; other times faults sole down onto the reflector (r) which would represent their basal detachment level.

7.2 DISCUSSION

Despite different hypothesis have been proposed in the past (e.g., [Artoni et al., 1992](#); [Storti, 1995](#); [Camurri, 2000](#); [Camurri et al., 2001](#); [Argnani et al., 2003](#); [Brozzetti et al., 2007](#)), the subsurface structures of the Northern Apennines are still debated. Proposing a comprehensive and fully satisfactory interpretation is difficult, and several first-order issues still need to be addressed: i) the deep prosecution toward the NE of the regional reflector (r) is not constrained; ii) the distribution of present-day seismicity does not completely fit the structures imaged in the seismic profiles; iii) cross-cutting relationships between such reflector and the thrust faults characterizing the central sector of the chain are debatable. In particular, the uncertainty associated with the latter point is not only related to the natural variability of the geological structures, but also to the subsurface data used for geological interpretations. Indeed, different geophysical processing of the same seismic profile might favor or reject the different interpretations. Moreover, the depth location of earthquake hypocenters is affected by significant uncertainty.

Solving such long-debated issues is out-of the purpose of this Ph.D. Thesis, so the details of such problems will not be addressed. However, a first order structural framework can be extrapolated from the integration of available geological and geophysical data, and the geodynamic context. Such structural framework can be used as a key to understand the case studies presented in previous chapters. In particular, it can be proposed that the high-angle extensional fault systems characterizing the internal sector of the chain crosscut a complex older structure that includes thrusting, large-scale folding, and thinning by low-angle fault systems. Additionally, the timing of the thrusting associated with the front of the Tuscan succession (t₁) can be constrained to an intermediate tectonic phase: it postdates the activity of the low-angle extensional fault system and is postdated by high-angle extensional faulting. Such a simple framework allows to reconcile the geometries imaged by subsurface data with the field geology evidences (Val di Lima fault system and Tellaro Detachment).

7.3 CONCLUSIONS

The internal sector of the Northern Apennines thrust wedge is characterized by different structures in the shallower ($\sim < 5$ km) and deeper

($\sim > 5$ km) portions. The shallower portion, in particular, is characterized by older low-angle extensional fault systems cut by younger high-angle extensional fault systems.

A first phase of low-angle extensional faulting can be correlated, by a cause-and-effect relationship, with deeper compressive structures. Indeed, the activity of deep compressive structures likely caused an uplift of the internal portion of the thrust wedge and possibly produced overcritical taper conditions. Such mechanical instability might have triggered extensional deformations in the shallower portion of the wedge, leading to the development of Late Miocene low-angle extensional fault systems such as the Val di Lima fault system (chapter 5). The conceptual model of an accretionary wedge characterized by remarkable strain partitioning (chapter 3) might be the key to understand this tectonic phase. According to this interpretation, the regional seismic reflector (r) characterizing the Northern Apennines subsurface structures might have constituted the roof thrust of a regional duplex structure resulted from basal accretion processes and antiformal stacking. Successively, the activity of low-angle fault systems would have continued in a late-orogenic framework, causing further thinning of the internal sector of the belt, as in the case of the Tellaro Detachment (chapter 6).

Eventually, the deformation front migrated toward the foreland and the internal sector of the Apenninic belt became affected by crustal extension in a back-arc setting. Since Pliocene times, high-angle extensional fault systems developed, associated with the opening of Lunigiana and Vara valley basins. The present-day seismotectonic framework of the Northern Apennines indicates that such high-angle faults are still active, and the hypocentral earthquake distribution suggests their prosecution at depths as great as 10 km. Consequently, it can be proposed that such extensional structures cut, or reactivate, the compressional structures observed below the regional seismic reflector (r), even if seismic imaging does not clearly support this interpretation.

GENERAL CONCLUSIONS

The overprinting of extensional and contractional deformations during the development of orogenic wedges has been studied through the analysis of analogue models, seismic profiles, and field-based investigation of selected case studies.

Analogue models allowed investigating few basic processes that affect the long-term wedge evolution. The influence of key parameters has been tested, and the mutual interplay between accretionary and erosive processes highlighted. Rheological layering of the subducting plate proved to exert a major control on mechanical heterogeneity and deformation partitioning, allowing the development of domains undergoing horizontal and vertical accretion. A strong feedback is observed between accretionary processes and material transfer induced by erosive processes. Experiments show a cyclical behavior of basal accretion, leading to episodic underplating of tectonic units, which has a strong impact on the vertical component of displacement of rock material. In turn, this changes the surface slope, favoring erosion in domains of strong surface uplift. Extensional deformations are not directly investigated in the models, due to limitation of the experimental methodology. Nevertheless, the experimental program allowed the definition of a large-scale mechanical framework to better understand the case studies presented in successive chapters.

After the characterization of orogenic wedges and the basic processes influencing their evolution, sin-contractional extensional deformations have been investigated through the analysis of selected case studies from two different orogens: the Oman Mountains (Sultanate of Oman) and the Northern Apennines (Italy). Selected examples include structures at different scales, developed at different structural levels. Moreover, they are located at the front, in the central, and in the internal portions of the wedges, allowing the investigation of structures associated to incipient and more evolved orogenic phases.

In the Jabal Qusaybah anticline, interactions between extensional and contractional deformations are investigated at the local scale. Research is performed with a multidisciplinary approach that integrates structural and petrographical-geochemical study of fault- and fold-related deformation structures. Multidisciplinary analysis indicates that folding and faulting occurred coevally during the growth of the anticline. Indeed, in the proposed evolutionary model, positive fault inversion, fold amplification, and strike-slip and extensional faulting mutually interacted. Sin-contractional extensional faults, in particular, are associated with along- and across-strike fold-related outer-arc extension or develop in transtensional horse-tails at strike-slip fault

zone tips. This work illustrates the possible complexity of deformation patterns in fault-related anticlines, with the presence of abundant extensional faults that is not predicted by conventional templates for folding-related fracturing. Similar atypical fault and fold patterns could be expected in structures at the toe of thrust wedges, particularly where evaporites provide the basal décollement. Consequently, when dealing with fracture pattern predictions for comparable subsurface structures, ignoring sin-contractional extensional deformations can lead to large errors of possible economic significance.

In the Val di Lima fault system case study, the interactions between extensional and contractional deformations have been investigated at the scale of a whole wedge sector. This case study highlights both the cause-and-effect and overprinting relationship between compressional and extensional deformations. Indeed, thrusting and antiformal stacking at depth caused wedge thickening and resulted in overcritical taper conditions that triggered extensional deformations at shallower depths in this sector of the belt. Additionally, the complex structural architecture clearly indicates that this area was cyclically affected by compressive and extensional structures, leading to the successive development of: i) a recumbent isoclinal fold; ii) a low-angle extensional fault system in the backlimb; iii) a system of NE-vergent open folds which locally verticalize the low-angle fault system; iv) late-tectonic high-angle faulting.

The Tellaro Detachment is located on the western side of the Punta Bianca promontory, in the most internal portion of the Apennines wedge. This area is characterized by an extremely thinned structural edifice, in which exhumed metamorphic units are exposed. As a consequence, low-angle extensional fault systems are best developed and less affected by superimposed compressional deformations. The Punta Bianca promontory, in particular, is thinned by low-angle extensional fault systems that dismember the previously stacked tectonostratigraphic units, and are then affected by superimposed high-angle faulting. Our research focused on the Tellaro Detachment, which is the best exposed among the low-angle extensional fault systems; appropriate exposure allowed us to better characterize the long-lasting fault activity during progressive exhumation of the Tuscan succession and the associated paleofluid evolution.

As a general conclusion, compressional and extensional deformations affecting orogenic wedges are investigated in this Ph.D. Thesis with a multidisciplinary approach, allowing to: i) understand the cause-and-effect relationship between them; ii) highlight the complexity of the deformation patterns developed due to the mutual overprinting of compressional and extensional structures; iii) characterize the permeability associated to the extensional fault zones and the effects on fluid circulation through the upper crust; iv) discuss the implications of sin-contractional extensional deformations on the me-

chanical behavior of thrust wedged and on the long-lasting evolution of orogens.

Additionally, research activities partly focused on low-angle extensional fault systems, contributing to the long-lasting debate whether extensional detachment faults can be activated at very shallow dip, or result from post-kinematic passive rotation of former high-angle fault zones. In particular, the optimal exposures of the Tellaro Detachment allowed to fully characterize the structural architecture and the paleofluid evolution of such a misoriented fault system. Our findings demonstrate that major low-angle detachments and subsidiary high-angle faults can be contemporaneously active in a fluid-assisted environment. Our results contribute to further support the notion that continental crustal thinning and faulting can occur also for shallow values of fault dip.

The results of this Ph.D. Thesis have scientific and industrial implications. Presented observations gives insights to discuss different processes such as: i) the exhumation of high-pressure metamorphic rocks in orogenic belts; ii) the activity of misoriented faults; iii) the effective exploitation of hydrocarbon reservoir through a better prediction of fracture permeability in faulted anticlines.

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