EDITED BY

Cathy Busby and **Antonio Azor**

TECTONICS OF SEDIMENTARY BASINS

RECENT ADVANCES

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WILEY-BLACKWELL

A John Wiley & Sons, Ltd., Publication

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Preface

The plate tectonic revolution of the 1960's provided the first unified framework for models on the origin of mountain belts and basins; this resulted in an outpouring of landmark papers in the 1970's and 1980's. When Ray Ingersoll and Cathy Busby taught Tectonics of Sedimentary Basins in the late 1980s (at UCLA and UCSB respectively), they were frustrated by a lack of textbooks or summary papers on this topic. Instead, professors were forced to compile impossibly long reading lists for their students, and try to synthesize the material for them. For this reason, Professors Busby and Ingersoll decided to edit a textbook on the topic for Blackwell, to be aimed at the senior undergraduate to professional geologist level. This was an exhaustive treatment that took five years to produce, and it was published in 1995 (Busby and Ingersoll, 1995). Online access to the 1995 book is provided by the publisher, because it still provides a valid and complete introduction to the topic. We recommend that the undergraduate geology student begin with the 1995 book, and that the graduate student and professional refer to it as needed while reading the new book.

Fifteen years later, there have been many advances in our understanding of the plate tectonic controls on basin formation and evolution. One large area of growth has been in the field of active tectonics, where advances in global positioning and stratigraphic or surface dating techniques allow workers to compare present-day plate motions with the growth of structures on the time scale of thousands of years. Our understanding of the sedimentary response to tectonic events has been improved through numerical/ analog modeling and detailed field observations. Major advances have been made in studies of the subsurface, through seismic surveys of crustal to upper mantle structure, as well as 3D seismic surveys of basin fills, in some areas augmented by cores. Isotopic studies of detrital minerals (e.g., U-Pb zircon) are now widely used to reconstruct tectonic events, including large-scale basin translation, patterns of unroofing in regions around basins, and reconstruction of

sediment pathways across continents through time. Paleomagnetic methods are now very widely employed for precise dating and correlation of strata, for determining sources and flow paths of widespread volcanic units, and for evaluating the importance of tectonic rotation. Another new approach to understanding tectonic problems is the use of ArcGIS to manipulate geochronological, geochemical, biostratigraphic, and paleomagnetic databases, in concert with satellite, air photograph, and geologic map data. At the same time, proliferation of detailed models for complex volcanic-volcaniclastic dispersal-depositional systems has permitted detailed tectonic reconstruction of a wider range of basin types, with dateable fill. Last but not least, numerical and analog modeling of geodynamic processes is more sophisticated than ever.

The new book presented here was produced in response to the demand for an update on the topic, hence the title *Tectonics of Sedimentary Basins: Recent Advances*. Our mission was to assemble an all-star cast in the field of basin tectonics, and give them a venue to present "cutting-edge" material. Unlike the 1995 book, this is not a comprehensive treatment of the entire subject; that was useful 15 years ago, but online access to publications has reduced the need for such comprehensive treatment. Instead, we think our new book represents the state of the art in research on basin tectonics. This book was produced very rapidly (in about two years), because it consist of numerous short chapters, most of them authored by only one or two people. The authors accomplished brevity by focusing on results of global importance and key issues raised over the last 15 years. Primary data are not generally presented here, but primary data sources are well cited, so the reader has a guide to all of the important recent literature on each topic. We believe the resulting chapters showcase methods and results that use innovative approaches and have led to a new understanding of tectonic processes, being, furthermore, "transportable" to other regions. We offer a very brief summary of

the main topics addressed in the book, giving the rationale for their order of appearance.

PART 1: "INTRODUCTION"

The first part of the book is an overview of the topic of this book, by Ingersoll (Chapter 1). This chapter is a thorough update on his grand overviews published in the Geological Society of America Bulletin in 1988, and in *Tectonics of Sedimentary Basins* (with Busby) in 1995. This chapter provides the reader with an understanding of the processes and nomenclature common to all succeeding chapters in the book.

PART 2: "NEW TECHNIQUES AND MODELING"

Everyone involved agreed that *Tectonics of Sedimentary Basins: Recent Advances* needed a part on techniques, because they have proliferated and been refined so much in the past 15 years (the 1995 book did not have a part like this). We made a conscious choice to limit discussions of applications in these chapters, thereby keeping each chapter as short as possible while providing references to applied studies. This part gives the reader an overview of the most important advances in techniques applied to tectonic analysis of sedimentary basins. Detrital zircon geochronologic techniques (Chapter 2) and terrestrial cosmogenic nuclide techniques (Chapter 3) have come into widespread use. Meanwhile, magnetostratigraphic techniques, seismic interpretation techniques, and basin/stratigraphic modeling techniques, while not entirely new, have become far more sophisticated (Chapters 4–8). This part of the book is missing a chapter on the huge and diverse field of chemostratigraphic techniques, which have been evolving for decades, but a treatment of that topic would require a second book.

The third and fourth parts of this book are organized by tectonic setting, divided into broadly divergent and broadly convergent margins.

PART 3: "RIFT, POST-RIFT, TRANSTENSIONAL, AND STRIKE-SLIP BASIN SETTINGS"

Part 3 opens with a chapter on the classic active orthogonal rift in East Africa (Chapter 9). It then moves to transtensional rift basins in a "successful" continental rift (Gulf of California, Chapter 10), and a transtensional rift that is still in progress (Walker Lane, Chapter 11). Transform margins are not dealt with here; for a global catalogue and description of strike-slip fault systems, see the 142-page opus by Paul Mann, published in 2007 (Geological Society of London Special Publications, vol. 290). In keeping with our theme of tectonics and sedimentation, Part 3 of our book also includes the so-called passive margins that show evidence of deformation long after sea-floor spreading began (Chapters 12 and 13).

PART 4: "CONVERGENT MARGINS"

Part 4 is broadly organized to move from the trench to more inboard settings, and from sea-floor subduction settings to collisional settings. This begins with an examination of processes involved in subduction initiation (Chapter 14), and continues with new results from what is probably the best-studied modern subduction complex on Earth (Chapter 15). Part 4 then proceeds through a study of forearc deformation by flat-slab subduction (Chapter 16), and the basinal record of bringing arcs into continental subduction zones (Chapter 17). It continues on the theme of sea-floor subduction by examining an Andean forearc basin (Chapter 18), as well as extensional and transtensional intra-arc basins of the Southwestern USA (Chapter 19). An overview of both subduction-related and collisional foreland basins follows (Chapter 20). Then we look at basins that lie on top of orogens, referred to as "hinterland basins" (a term included in the revised nomenclature of Chapter 1). These are described from both subduction and collisional settings (Andes and Tibet), on the time scale from millions of years (Chapter 21) to hundreds of thousands and thousands of years (Chapter 22). We then move to "intramontane" basins of the Betic Cordillera of Spain, interpreted to be extensional basins formed in a late orogenic setting due to mantle delamination and/or slab rollback or detachment (Chapter 23). Part 4 then moves inboard in tectonic setting, to the foreland. Chapter 24 examines patterns of flexural subsidence in the western Canada foreland basin, inferred to be broadly controlled by oceanic plate subduction, and at the Cordilleran scale controlled by terrane accretion events. Chapter 25 contrasts the elements of a typical contiguous foreland basin (Bolivia)

with those of a broken foreland (Argentina). These case studies are followed by studies that deal with general kinematic models for thrust wedge-foreland systems (Chapter 26), and models for growth fault-related folds in contractional settings (Chapter 27). Part 4 has an unfortunate absence of oceanic/island arc convergent margin basin tectonic studies; that, too, is deserving of a separate book.

PART 5: "PLATE INTERIOR BASINS AND WIDESPREAD BASIN TYPES"

The last part of the book treats sedimentary basin tectonic topics that do not fit neatly into divergent or convergent plate tectonic settings. Plate interior poly-phase (PIP) basins are important for their size and long-term structural and stratigraphic record (Chapter 28). The vast sedimentary record of the Grenvillian tectonomagmatic event is described in the context of supercontinent assembly (Chapter 29). In Chapter 30, cratonic basins are described as long-lived circular or elliptical crustal sags on thick, relatively stable continental lithosphere, and are interpreted to be primarily formed by protracted plate-wide stretching at low strain rate. Last, Chapter 31 describes the distinctive stratigraphic and sedimentary facies characteristics that are common to endorheic (internally drained) basins in a wide variety of tectonic settings.

There were at least a few additional topics that were garnering increasing attention at the time this book was being produced. The halokinetic basin, important for commonly containing petroleum, is now included in Ingersoll's revised nomenclature (Chapter 1), and it deserved its own chapter here. New techniques are rapidly evolving to deduce paleo-elevations from sedimentary basin fills, using stable isotopes in paleosols, fossils, silicates, and volcanic glasses, but these are treated elsewhere (e.g., see Reviews in Mineralogy & Geochemistry, 2007, vol. 66). A very rapidly expanding, huge field of research uses global seismic tomography studies to infer linkages between mantle and surface processes. For example, how important is the role of "mantle/lithospheric drips" in causing surface subsidence (e.g., the Tulare Lake Basin of the San Joaquin Valley, California)? Are they too small and too transient to be important? How does the subduction of huge oceanic plateaus control uplift and subsidence events on continents?

And how do surface tectonic basin processes provide a record of tomographically imaged subduction processes? Slab rollback is discussed in this book, but what are the effects of other processes, such as stalled slabs and broken slabs?

Like all geologic research, this book is a "progress report" on our understanding of the tectonics of sedimentary basins, which we think has advanced greatly in recent years. We hope you find the book useful. We give our sincere thanks to the many reviewers who worked hard to give us valuable feedback (listed below).

Cathy Busby and Antonio Azor, Editors
February 12, 2011

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Part 1

Introduction

Chapter 1

Tectonics of sedimentary basins, with revised nomenclature

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ABSTRACT

Actualistic plate-tectonic models are the best framework within which to understand the tectonics of sedimentary basins. Sedimentary basins develop in divergent, intraplate, convergent, transform, hybrid, and miscellaneous settings. Within each setting are several variants, dependent on type(s) of underlying crust, structural position, sediment supply, and inheritance. Subsidence of sedimentary basins results from (1) thinning of crust (2) thickening of mantle lithosphere (3) sedimentary and volcanic loading (4) tectonic loading (5) subcrustal loading (6) asthenospheric flow, and (7) crustal densification. Basins vary greatly in size, life span, and preservation potential, with short-lived basins formed in active tectonic settings, especially on oceanic crust, having low preservation potential, and long-lived basins formed in intraplate settings having the highest preservation potential.

Continental rifts may evolve into nascent ocean basins, which commonly evolve into active ocean basins bordered by intraplate continental margins with three types of configurations: shelf-slope-rise, transform, and embankment. Continental rifts that do not evolve into oceans become fossil rifts, which later become sites for development of intracratonic basins and aulacogens. If all plate boundaries within and around an ocean basin become inactive, a dormant ocean basin develops, underlain by oceanic crust and surrounded by continental crust.

Sites for sedimentary basins in convergent settings include trenches, trench slopes, forearcs, intra-arcs, backarcs, and retroarcs. Complex dynamic behavior of arc-trench systems results in diverse configurations for arc-related basins. Most notable is the overall stress regime of the arc-trench system, with resulting response along and behind the magmatic arc. Intra-arc rifting in highly extensional arcs commonly evolves into backarc spreading to form new oceanic crust. Backarcs of neutral arcs can contain any type of preexisting crust that was trapped there at the time of initiation of the related subduction zone. Highly compressional arcs develop retroarc foldthrust belts and related retroforeland basins, and may develop hinterland basins; in extreme cases, broken retroforelands may develop in former cratonal areas.

As nonsubductable continental or arc crust is carried toward a subduction zone, collision generally initiates at one point and the resulting suture propagates away from this point of initial impact. Remnant ocean basins form on both sides of the initial impact point, and rapidly fill with sediment derived from the suture zone. As collision continues, the flux of sediment into the remnant ocean basin(s) increases concurrently with shrinkage of the basin until final suturing and obduction of the accreted sediment occur. Concurrently with collision, proforeland basins form on continental crust of the subducting plate and collisional retroforeland basins form on the overriding plate. Impactogens, broken forelands, and hinterland basins also may result.

In transform settings and along complex strike-slip fault systems related to convergent settings, changing stress regimes related to irregularities in fault trends, rock types, and plate motions result in transtension, transpression, and transrotation, with associated complex, diverse, and short-lived sedimentary basins.

Two previously unnamed basin types that have received increasing attention recently are halokinetic basins (related to salt tectonics, especially along intraplate margins with embankment configurations) and bolide basins (resulting from extraterrestrial impacts). Sediment accumulates in successor basins following cessation of basin-controlling processes, whether in divergent, convergent, transform, or hybrid settings.

The ultimate goal of classifying and reviewing all types of sedimentary basins is the improvement of paleotectonic and paleogeographic reconstructions through the application of actualistic models for basin evolution. Interdisciplinary studies that test and refine these models will improve our knowledge of Earth history.

Keywords: basin nomenclature; plate-tectonic settings; subsidence mechanisms; preservation potential; paleotectonic reconstruction

INTRODUCTION

It has been more than a decade since I reviewed and revised my original basin classification (i.e., Ingersoll, 1988; Ingersoll and Busby, 1995), which was based primarily on Dickinson's (1974b, 1976a) statement of fundamental principles that should guide discussion of the tectonics of sedimentary basins. Many new insights and models have been developed recently; in addition, nomenclature has evolved in complex ways. Therefore, now is an appropriate time to consolidate, revise, and discuss how to communicate about the tectonics of sedimentary basins.

As in my previous papers on this subject, I follow Dickinson's (1974b, 1976a) suggestions that nomenclature and classification be based on the following actualistic plate-tectonic processes and characteristics, which ultimately control the location, initiation, and evolution of sedimentary basins in diverse tectonic settings. Horizontal motions of plates, thermal changes through time, stretching and shortening of crust, isostatic adjustments, mantle dynamics, surficial processes, and even extraterrestrial events influence sedimentary basins. Additional study of sedimentary basins, inevitably, leads to greater complexity of models to explain them. Although we should search for unifying principles that lead to deeper understanding of processes and results, the complexity of the real world dictates that enhanced knowledge about sedimentary basins results in more complex models. Thus, new types of sedimentary basins are added to the list provided in Ingersoll and Busby (1995) because these are actual features that need to be understood. Gould (1989, 98) stated, "Classifications are theories about the basis of natural order, not dull catalogues compiled only to avoid chaos." I hope that my discussion serves the dual purposes of reducing nomenclatural chaos and suggesting a framework within which to understand the complex controls on the origin and evolution of sedimentary basins.

NOMENCLATURE

First-order criteria for classifying sedimentary basins (Dickinson, 1974b, 1976a) are (1) type of nearest plate boundary(ies) (2) proximity of plate boundary(ies), and (3) type of substratum. Thus, the first-order classification, based on criteria (1) and (2) is divergent, intraplate, convergent, transform, hybrid, and miscellaneous settings (Table 1.1). Within each of these categories are several variants, dependent on type of substratum (oceanic, transitional, continental, and anomalous crust), as well as structural position, sediment supply, and inheritance.

Basin classification and nomenclature are based on characteristics of a basin *at the time of sedimentation*. Thus, many stratigraphic successions are multidimensional and multigenerational in terms of plate-tectonic controls on their evolution. A single stratigraphic succession may represent several different tectonic settings. "The evolution of a sedimentary basin thus can be viewed as the result of a succession of discrete plate-tectonic settings and plate interactions whose effects blend into a continuum of development" (Dickinson, 1974b, 1).

It is important to realize that "basin," as used herein, refers to any stratigraphic accumulation of sedimentary or volcanic rock; the three-dimensional architecture of basins may approximate saucers, wedges, sheets, and odd shapes.

Table 1.1. Basin classification with modern and ancient examples

Setting	Basin type	Definition	Modern example	Ancient example	Analog model (Figure)
Divergent	Continental rifts	Rifts within continental crust, commonly associated with bimodal magmatism	Rio Grande rift	Proterozoic Keweenawan rift	3B
	Nascent ocean basins and continental margins	Incipient ocean basins floored by new oceanic crust and flanked by young rifted continental margins	Red Sea	Jurassic of East Greenland	3C
Intraplate	Intraplate continental margins				
	Shelf-slope-rise configuration	Mature rifted intraplate continental margins with shelf edge near boundary between continental and oceanic crust	East Coast of USA	Early Paleozoic of USA and Canadian Cordillera	3D
	Transform configuration	Intraplate continental margins that originate along transform plate boundaries	South Coast of West Africa	Precambrian-early Paleozoic Alabama-Oklahoma transform	3E
	Embankment configuration	Progradational intraplate continental margins with shelf edge above oceanic crust	Mississippi River Gulf Coast	Early Paleozoic Meguma terrane of Canadian Appalachians (?)	3F
	Intracratonic basins	Broad cratonic basins underlain by fossil rifts	Chad basin	Paleozoic Michigan basin	3A
	Continental platforms	Stable cratons with thin and laterally extensive sedimentary strata	Barents Sea	Middle Paleozoic, North American midcontinent	3A
	Active ocean basins	Basins floored by oceanic crust formed at active divergent plate boundaries unrelated to arc-trench systems	Pacific Ocean	Various ophiolitic complexes (?)	3G
	Oceanic islands, seamounts, aseismic ridges, and plateaus	Sedimentary aprons and platforms formed in intraoceanic settings other than arc-trench systems	Emperor-Hawaii seamounts	Mesozoic Snow Mountain Volcanic Complex (Franciscan of northern California)	3G
	Dormant ocean basins	Basins floored by oceanic crust, which is neither spreading nor subducting	Gulf of Mexico	Paleozoic Tarim basin (China) (?)	3H

(*continued*)

Table 1.1. (*Continued*)

Setting	Basin type	Definition	Modern example	Ancient example	Analog model (Figure)
Convergent	Trenches	Deep troughs formed at oceanic subduction zones	Chile Trench	Cretaceous, Shumagin Island (southern Alaska)	4A
	Trench-slope basins	Local structural depressions on subduction complexes	Central American Trench	Cretaceous Cambria slab (central California)	4B
	Forearc basins	Basins within arc-trench gaps	Offshore Sumatra	Cretaceous Great Valley (California)	4B
	Intraarc basins *Oceanic intraarc basins*	Basins along intraoceanic arc platforms, which include superposed and overlapping volcanoes	Izu Bonin arc	Copper Hill, Gopher Ridge Complex (Jurassic, California)	4A
	Continental intraarc basins	Basins along continental-margin arc platforms, which include superposed and overlapping volcanoes	Lago de Nicaragua	Early Jurassic Sierra Nevada (eastern California)	4C
	Backarc basins *Oceanic backarc basins*	Oceanic basins behind intraoceanic magmatic arcs (including interarc basins between active and remnant arcs)	Marianas backarc	Jurassic Josephine ophiolite (northern California)	4A,B
	Continental backarc basins	Continental basins behind continental-margin arcs without foreland foldthrust belts	Sunda Shelf	Late Triassic - Early Jurassic of USA Cordillera	4C
	Retroforeland basins *Retroarc foreland basins*	Foreland basins on continental sides of continental-margin arc-trench systems	Andes foothills	Cretaceous Sevier foreland of USA Cordillera	4E
	Collisional retroforeland basins	Foreland basins formed on overriding plates during continental collisions (may have retroarc precursors)	Western Tarim basin (China)	Triassic-Jurassic Ordos basin (China)	4F
	Broken-retroforeland basins	Basins formed among basement-cored uplifts in retroforeland setting	Sierras Pampeanas basins (Argentina)	Late Cretaceous-Paleogene Laramide basins of USA Cordillera	4D
	Remnant ocean basins	Shrinking ocean basins between colliding continental margins and/or arc-trench systems, and ultimately subducted or deformed within suture belts	Bay of Bengal	Pennsylvanian-Permian Ouachita basin	4E

	Basin type	Description	Example 1	Example 2	Code
	Proforeland basins	Foreland basins formed on continental crust that is part of the subducting plate during continental and/or arc collision	Persian Gulf	Mid-Cenozoic Swiss Molasse basin	4F
	Wedgetop basins	Basins formed and carried on moving thrust sheets	Peshawar basin (Pakistan)	Neogene, Apennines (Italy)	4F
	Hinterland basins	Basins formed on thickened continental crust behind foreland foldthrust belts	Altiplano Plateau (Bolivia)	Neogene Zhada basin (Tibet)	4D
Transform	Transtensional basins	Basins formed by extension along strike-slip releasing bends and steps	Dead Sea	Carboniferous Magdalen basin (Gulf of Saint Lawrence)	5A
	Transpressional basins	Basins formed by shortening along strike-slip constraining bends and steps	Santa Barbara Basin (foreland type) (California)	Miocene Ridge basin (fault-bend type) (California)	5B
	Transrotational basins	Basins formed by rotation of crustal blocks about vertical axes within strike-slip fault systems	Western Aleutian forearc (?)	Miocene Los Angeles basin (Topanga basin) (California)	5C
Miscellaneous and hybrid	Aulacogens	Reactivated fossil rifts at high angles to orogenic belts	Mississippi embayment	Paleozoic Anadarko aulacogen (Oklahoma)	6A
	Impactogens	Newly formed continental rifts at high angles to orogenic belts, without preorogenic history (in contrast to aulacogens)	Baikal rift (distal) (Siberia)	Rhine Graben (proximal) (Europe)	6B
	Collisional broken foreland	Diverse basins formed on deformed continental crust due to distant collisions	Qaidam basin (China)	Pennsylvanian-Permian Ancestral Rocky Mountain basins of the USA Cordillera	6B
	Halokinetic basins	Basins formed due to deformation of salt, most commonly in continental embankments and proforelands	Mini-basins of deep Gulf of Mexico	Cretaceous-Paleogene La Popa basin (Mexico)	3F
	Bolide basins	Depressions in Earth's surface resulting from extraterrestrial impacts	Meteor Crater (Arizona)	Cretaceous-Paleogene Chicxulub basin (Mexico)	3E
	Successor basins	Basins formed in intermontane settings following cessation of local taphrogenic or orogenic activity	Southern Basin and Range (Arizona)	Paleogene Sustut basin (?) (British Columbia)	5C

Modified after Ingersoll and Busby (1995).

Also, basins may form by subsidence of a substrate, development of a barrier to transport of sediment, filling of a preexisting hole, or relative movement of source and sink.

SUBSIDENCE MECHANISMS AND PRESERVATION POTENTIAL

Surfaces of deposition may subside due to the following processes (Dickinson, 1974b, 1976a, 1993; Ingersoll and Busby, 1995) (Table 1.2): (1) thinning of crust due to stretching, erosion, and magmatic withdrawal (2) thickening of mantle lithosphere during cooling (3) sedimentary and volcanic loading (local crustal isostasy or regional lithospheric flexure) (4) tectonic loading of both crust and lithosphere (5) subcrustal loading of both crust and lithosphere (6) dynamic effects of asthenospheric flow, and (7) crustal densification. Figure 1.1 illustrates that crustal thinning dominates during early stages of extension (e.g., rifts and transtensional basins), and mantle-lithospheric thickening dominates following the initiation of seafloor spreading (during the rift-to-drift transition along divergent margins which evolve into intraplate margins). Sedimentary loading dominates along continental-oceanic crustal boundaries which are supplied by major rivers and deltas (e.g., continental embankments and remnant ocean basins). Tectonic loading dominates in settings where crustal shortening dominates (e.g., trenches and foreland basins). The other three types of subsidence mechanisms are generally subordinate.

The diversity of tectonic and structural settings of sedimentary basins dictates that they vary greatly in size, life span, and preservation potential (Fig. 1.2) (Ingersoll, 1988; Ingersoll and Busby, 1995; Woodcock, 2004). Many sediment accumulations are destined to be destroyed relatively soon after deposition (e.g., most basins residing on oceanic crust or in rapidly uplifting orogenic settings). In contrast, basins formed during and following stretching of continental crust (e.g., continental rifts that either evolve into seafloor spreading or fail to do so) have high preservation potential because they subside and are buried beneath intraplate deposits following rifting. On the other hand, stratigraphic sequences along intraplate continental margins are destined to be partially subducted as they are pulled into trenches, thus preserving them at moderate to deep crustal levels as highly deformed and metamorphosed terranes. Such metasedimentary and metavolcanic terranes, along with voluminous sediments deposited in remnant ocean basins, are major rock bodies involved in the construction of continental crust, although their substrates (oceanic crust) are mostly subducted (e.g., Graham et al., 1975; Ingersoll et al., 1995, 2003).

DIVERGENT SETTINGS

Sequential rift development and continental separation

The relative importance of "active" (mantle-convective-driven) versus "passive" (lithospheric-driven) processes during initiation of continental rifting is debated (e.g., Sengor and Burke, 1978; Ingersoll and Busby, 1995; Sengor, 1995). Regardless of the mechanisms of initiation of rifting, continental rifts may experience two life paths: "successful" rifting that evolves into seafloor spreading to form nascent ocean basins (Ingersoll and Busby, 1995; Leeder, 1995), which then evolve into active ocean basins with paired intraplate margins (Fig. 1.3), or "failed" rifting, which does not evolve into nascent ocean basins,

Table 1.2. Subsidence mechanisms

Crustal thinning	Extensional stretching, erosion during uplift, and magmatic withdrawal
Mantle-lithospheric thickening	Conversion of asthenosphere to mantle lithosphere during cooling following cessation of stretching and/or heating
Sedimentary and volcanic loading	Local isostatic compensation of crust and/or regional lithospheric flexure during sedimentation and volcanism
Tectonic loading	Local isostatic compensation of crust and/or regional lithospheric flexure during overthrusting and/or underpulling
Subcrustal loading	Lithospheric flexure during underplating of dense lithosphere
Asthenospheric flow	Dynamic effects of asthenospheric flow, commonly due to descent or delamination of subducted lithosphere
Crustal densification	Increased density of crust due to changing pressure/temperature conditions and/or emplacement of higher-density melts into lower-density crust

Fig. 1.1. Suggested subsidence mechanisms for all types of sedimentary basins.

instead producing fossil rifts, commonly overlain by intracratonic basins (Sengor, 1995). Ingersoll and Busby (1995), Leeder (1995), and Sengor (1995) reviewed most aspects of continental stretching, basin formation, structural development, and different life paths during and after continental rifting. Here, I highlight changes in nomenclature and models involved in the evolution from continental rifts to intraplate margins (the rift-drift transition).

Fig. 1.2. Typical life spans for sedimentary basins versus their post-sedimentation preservation potential, which refers to average time interval during which basins will not be uplifted and eroded during and following sedimentation. Sedimentary or volcanic fill may be preserved as accretionary complexes during and after basin destruction (especially true for all strata deposited on oceanic crust). Intraplate continental margins have high preservation potential in the sense of retaining their basement, but are likely to be highly deformed and metamorphosed beneath and within suture belts, and may be difficult to recognize in ancient settings.

Fig. 1.3. True-scale actualistic analog models for sedimentary basins in divergent, intraplate and miscellaneous settings. Mantle lithosphere thins during decompression melting as plates diverge; mantle lithosphere thickens during cooling, following cessation of divergence. Also shown are two miscellaneous basins (bolide and halokinetic). Placement of bolide basin is arbitrary; they may form anywhere on Earth's surface, although preservation is more likely in cratonal areas (as shown in E). Halokinetic basins may form anywhere that salt is deeply buried; however, continental embankments (as shown in F) are the most common locations. Continental crust = jackstraw pattern; oceanic crust = vertical lines; mantle lithosphere and derived igneous rocks = black; asthenosphere and derived melts = orange; salt (halokinetic only) = black.

Continental rifts

The most common basins associated with continental rifts (Fig. 1.3b) ("terrestrial rift valleys" of Dickinson, 1974b; Ingersoll, 1988) are half grabens developed on the hanging walls of normal faults (Leeder and Gawthorpe, 1987; Leeder, 1995; Gawthorpe and Leeder, 2000). Gawthorpe and Leeder (2000) summarized conceptual models for the tectono-sedimentary evolution of continental rift basins, including their three-dimensional development. They discussed structural, geomorphic, climatic, and lake/sea-level influences on basin development.

All of the models presented by Gawthorpe and Leeder (2000) involve high-angle normal faults. In these half grabens, most sediment is derived from the hanging wall, whereas the coarsest material, which is derived primarily from the footwall, is restricted to small steep alluvial fans or fan deltas along the faulted basin boundary. In contrast, supradetachment basins (formed above low-angle normal faults) receive most of their sediment from the breakaway footwall and tend to be dominated by coarse-grained detritus (Friedmann and Burbank, 1995). Additional variants on the Gawthorpe and Leeder (2000) half-graben model include development of accommodation zones, relay ramps, anticlinal-full-graben basins, and synclinal-horst basins (Rosendahl, 1987; Faulds and Varga, 1998; Ingersoll, 2001; Mack et al., 2003).

Nascent ocean basins and continental margins

As continental lithosphere is stretched and thinned, mantle asthenosphere eventually rises close to the surface (Fig. 1.3c). During the transition from continental rifting to seafloor spreading, transitional crust forms, either as stretched continental crust (quasicontinental) or sediment-rich basaltic crust (quasioceanic) (Dickinson, 1974b; Ingersoll, 2008b). Continental rifting evolves into seafloor spreading only in the absence of significant sediment so that oceanic crust is the only solid material with which rising asthenospheric melts can interact (Einsele, 1985; Nicolas, 1985). Thus, a significant width of transitional crust typically forms on the margins of nascent ocean basins prior to initiation of true seafloor spreading.

As these transitional types of crust form and the two continental margins move apart, a nascent ocean basin develops ("proto-oceanic gulf" and "narrow ocean" of Dickinson (1974b);

"proto-oceanic rift trough" of Ingersoll (1988)). The Red Sea is the type nascent ocean basin, with active seafloor spreading, clastic and carbonate sedimentation along the margins, and uplifted rift shoulders along the continental margins (Cochran, 1983; Bohannon, 1986a, 1986b; Coleman, 1993; Leeder, 1995; Purser and Bosence, 1998; Bosworth et al., 2005). Thick evaporite deposits may form during the transition from rift basin to nascent ocean basin, as well as during much of the history of nascent ocean basins, given the right combination of arid climate, limited communication with other marine bodies, and lack of detrital input (Dickinson, 1974b). The Gulf of California is an example of a transtensional nascent ocean basin (e.g., Atwater, 1989; Lonsdale, 1991; Atwater and Stock, 1998; Axen and Fletcher, 1998).

INTRAPLATE SETTINGS

Intraplate continental margins

Nascent ocean basins evolve into wide (Atlantic-type) oceans as two continents diverge along spreading ridges. During this evolutionary process, the newly rifted continental margins with uplifted rift flanks cool and subside as they move away from the spreading ridge. This process is referred to as the "rift-to-drift" transition, as a divergent setting evolves into an intraplate setting (Dickinson, 1974b, 1976a; Ingersoll, 1988; Bond et al., 1995; Ingersoll and Busby, 1995). Withjack et al. (1998) discussed complications in timing and process during this transition.

Subsidence mechanisms evolve from (1) thinning of continental crust by stretching and erosion during doming and rifting, to (2) thermal subsidence following rifting as the intraplate margin moves away from the spreading ridge, to (3) both local crustal and regional lithospheric sediment loading during the later history of the intraplate continental margin (Bond et al., 1995; Ingersoll and Busby, 1995). Lower-crustal and subcrustal flow and densification can locally modify subsidence.

Shelf-slope-rise configuration

Most mature intraplate continental margins consist of a seaward thickening wedge of shelf deposits on top of continental crust, which is thinner seaward (Fig. 1.3d). Transitional crust (both quasicontinental and quasioceanic; Dickinson, 1974b,

1976a) underlies the seaward transition from thick shelf deposits to thin slope deposits, which, in turn, merge into thick turbiditic rise and abyssal-plain deposits on oceanic crust (Bond et al., 1995; Ingersoll and Busby, 1995). Most modern Atlantic continental margins have this configuration, with carbonate environments dominating at lower latitudes devoid of extensive clastic input.

Transform configuration

Intraplate continental margins that originate along transform boundaries rather than rift boundaries have narrower sediment prisms and transitional crust (Fig. 1.3e). Tens of millions of years may pass between the time of initiation of transform motion (coincident with the rift-to-drift transition on adjoining margins) and the time of intraplate sedimentation (following passage of the spreading ridge along the transform boundary) (e.g., Bond et al., 1995; Turner et al., 2003; Wilson et al., 2003). The southern coast of West Africa exemplifies these characteristics; the latest Proterozoic - early Paleozoic Alabama-Oklahoma transform margin is an ancient example (e.g., Thomas, 1991).

Embankment configuration

Major rivers along intraplate continental margins commonly are localized by fossil rifts trending at high angle to the margins (Burke and Dewey, 1973; Dickinson, 1974b; Audley-Charles et al., 1977; Ingersoll and Busby, 1995). The best examples are the Niger Delta (Burke, 1972) and the Mississippi Delta (Worrall and Snelson, 1989; Salvador, 1991; Galloway et al., 2000), where the shelf edge has prograded over oceanic crust because the maximum sediment thickness allowed by isostatic loading (16–18 km; Kinsman, 1975) has been reached inland of the shelf edge (Fig. 1.3f). In the case of the USA Gulf Coast, several rivers in addition to the Mississippi have contributed to considerable progradation of the continental margin over a wide area; this is the type example of a continental embankment, a distinctly different configuration than either the shelf-slope-rise or transform configuration.

Intracratonic basins

Most intracratonic basins (e.g., Michigan basin) overlie fossil rifts (e.g., DeRito et al., 1983; Quinlan, 1987; Klein, 1995; Sengor, 1995; Howell

and van der Pluijm, 1999) (Fig. 1.3a). Renewed periods of subsidence in cratonic basins can generally be correlated with changes in lithospheric stress related to orogenic activity in neighboring orogenic belts (DeRito et al., 1983; Howell and van der Pluijm, 1999). Subsidence occurs when lithospheric rigidity lessens, allowing uncompensated mass in the upper crust (remnants of fossil rifts) to subside over a broad area. Between times of orogenic activity, the lithosphere strengthens so that attainment of local isostatic equilibrium is interrupted. Thus, an intracratonic basin may take hundreds of millions of years to reach full isostatic compensation (DeRito et al., 1983; Ingersoll and Busby, 1995; Howell and van der Pluijm, 1999).

Continental platforms

Cratonal stratigraphic sequences primarily reflect global tectonic events and eustasy (e.g., Sloss, 1988; Bally, 1989), although mantle dynamics, and local and regional events also influence continental platforms (e.g., Cloetingh, 1988; Burgess and Gurnis, 1995; van der Pluijm et al., 1997; Burgess, 2008). In contrast to intracratonic basins, platforms (Fig. 1.3a) accumulate sediment of uniform thickness over continental scales. Platformal stratigraphic sequences are transitional into continental margins, intracratonic basins, foreland basins, and other tectonic settings along continental margins (Ingersoll and Busby, 1995; Burgess, 2008). The distinction of distal foreland and platform sequences may be arbitrary, especially during times of high sea level, high carbonate productivity, and broad foreland flexure. Eustatically induced cyclothems are best expressed on platforms (e.g., Heckel, 1984; Klein, 1992; Klein and Kupperman, 1992), and paleolatitude and paleoclimate signals are best isolated in platformal sequences (Berry and Wilkinson, 1994). Platforms have generally experienced exposure and erosion during times of supercontinents, and have experienced maximum flooding approximately 100 My after supercontinent breakup (Heller and Angevine, 1985; Cogne et al., 2006).

Active ocean basins

The systematic exponential thermal decay of oceanic lithosphere as it moves away from spreading ridges is expressed by increasing water depth with age of oceanic crust (Sclater et al., 1971; Parsons and Sclater, 1977; Stein and Stein, 1992)

(Fig. 1.3g). As oceanic crust subsides with age and distance from spreading ridges, systematic pelagic and hemipelagic deposits accumulate (Berger, 1973; Heezen et al., 1973; Winterer, 1973; Berger and Winterer, 1974). Carbonate ooze accumulates above the carbonate compensation depth (CCD), which is depressed under areas of high biologic productivity; silica ooze accumulates above the poorly defined silica compensation depth (SCD); and only abyssal clay accumulates below the SCD. The result is a dynamic and predictive stratigraphy relating the age, depth, and paleoaltitude of oceanic crust to oceanic depositional facies. Volcaniclastic and turbidite deposits near magmatic arcs and continental margins complicate predicted stratigraphic sequences on oceanic plates (e.g., Cook, 1975; Ingersoll and Busby, 1995).

Oceanic islands, seamounts, aseismic ridges, and plateaus

Islands, seamounts, ridges, and plateaus thermally subside as oceanic plates migrate away from spreading ridges. Thermal anomalies independent of spreading ridges (e.g., hot spots) create new islands, ridges, and plateaus, which may have complex subsidence histories, dependent on their magmatic histories. Clague (1981) divided the post-volcanic history of seamounts into three sequential stages: subaerial, shallow water, and deep water or bathyal (Ingersoll, 1988; Ingersoll and Busby, 1995). As an island is eroded and subsides, fringing reefs and atolls may form, depending on latitude, climate, and relative sea level (e.g., Jenkyns and Wilson, 1999; Dickinson, 2004). Oceanic features, which may become accreted terranes at convergent margins (e.g., Wrangellia of the North American Cordillera; Ricketts, 2008), range in size from small seamounts to large mafic igneous provinces, such as the Ontong Java Plateau and related features (e.g., Taylor, 2006).

Dormant ocean basins

Dormant ocean basins are floored by oceanic crust, which is neither spreading nor subducting; in other words, there are no active plate margins within or adjoining the basin (Ingersoll and Busby, 1995) (Fig. 1.3h). This setting contrasts with active ocean basins, which include at least one active spreading ridge (e.g., Atlantic, Pacific,

and Indian oceans), and remnant ocean basins, which are small shrinking oceans bounded by at least one subduction zone (e.g., Bay of Bengal and Huon Gulf). The term "dormant" implies that there is no orogenic or taphrogenic activity within or adjacent to the basin; "oceanic" requires that the basin is underlain by oceanic lithosphere, in contrast to intracratonic basins, which are typically underlain by partially rifted continental lithosphere (Ingersoll and Busby, 1995).

Dormant ocean basins are created by two contrasting processes: (1) spreading ridges of nascent ocean basins cease activity (e.g., Gulf of Mexico; Pindell and Dewey, 1982; Pindell, 1985; Dickinson and Lawton, 2001), or (2) backarc basins (either extensional or neutral) are not subducted during suturing of continents and/or arcs (e.g., Black Sea; Okay et al., 1994) or South Caspian basin (Brunet et al., 2003; Vincent et al., 2005). The origin of dormant ocean basins may be difficult to determine because basement and original strata commonly remain deeply buried for hundreds of millions of years following cessation of seafloor spreading (e.g., Tarim and Junggar basins of western China) (e.g., Sengor et al., 1996). Following cessation of plate activity within and around the basin, sediment loading is the dominant subsidence mechanism, although lithospheric thickening due to residual cooling may be important (Ingersoll and Busby, 1995). Dormant ocean basins may have life spans of hundreds of millions of years and may vary considerably in size. The modern Gulf of Mexico, the largest known dormant ocean basin, is filling rapidly along its northern margin (the continental embankment of the Gulf Coast), but still contains oceanic crust with thin sediment cover in the south (e.g., Buffler and Thomas, 1994; Galloway et al., 2000; Dickinson and Lawton, 2001). The South Caspian Basin is small and partially filled with sediment (locally over 20 km thick; Brunet et al., 2003), and yet still is an oceanic basin. In contrast, the Tarim basin has a comparable sediment thickness, but is completely filled. These three basins are likely underlain by oceanic crust, or in the case of Tarim, an oceanic Plateau (Sengor et al., 1996); their long histories of cooling means that they are also underlain by thick and strong mantle lithosphere (Ingersoll and Busby, 1995). When a dormant ocean basin is filled to sea level, it may superficially resemble an intracratonic basin. The former, however, contains 16–20 km of sedimentary strata on top of strong oceanic lithosphere, whereas the latter

contains a few km of sedimentary strata underlain primarily by continental crust, with one or more fossil rifts beneath the basin center. Thus, when in-plate stresses affect dormant ocean basins and their surroundings, deformation usually occurs along their weak boundaries, whereas deformation of intracratonic basins is concentrated along the fossil rifts underlying their interiors. Foreland basins may form above the edges of dormant ocean basins during contractional deformation (e.g., the margins of the modern Tarim basin). Intracratonic basins may experience renewed subsidence or inversion tectonics (e.g., the modern North Sea) (Cooper and Williams, 1989; Cameron et al., 1992).

CONVERGENT SETTINGS

Arc-trench systems

Arc-trench systems may be categorized into three fundamental types: (1) extensional (2) neutral, and (3) compressional (Dickinson and Seely, 1979; Dewey, 1980) (Fig. 1.4). Arc-trench systems with significant strike slip may be considered a fourth type (Dorobek, 2008); strike-slip faults may occur in all types of arc-trench system, but they are especially common in strongly coupled systems experiencing oblique convergence (Beck, 1983). Many parameters determine the behavior of arc-trench systems, but the most important factors appear to be (1) convergence rate (2) slab age, and (3) slab dip (Molnar and Atwater, 1978; Uyeda and Kanamori, 1979; Jarrard, 1986; Kanamori, 1986), based on analyses of modern arc-trench systems (although see Cruciani et al., 2005, for an alternative interpretation). A major question arises from these analyses of contemporary Earth: is the present arrangement of spreading ridges and arc-trench systems typical of Earth history or an unusual configuration? Almost all modern east-facing arcs (e.g., Marianas) are extensional, with subduction of old lithosphere at steep angles. Almost all west-facing arcs (e.g., Andes) are compressional, with subduction of young lithosphere at shallow angles. Most south-facing arcs (e.g., Aleutians) are neutral, with subduction of middle-aged lithosphere at moderate angles. There are no north-facing arcs. Thus, it is very difficult to separate the covarying parameters of slab age, slab dip, facing direction, and type of arc-trench system. There is growing consensus (although see Schellart, 2007, 2008, for a contrary view) that facing direction of arc-trench systems

may be the fundamental determinant of the behavior of arc-trench systems because of west-ward tidal lag of the eastward rotating planet (e.g., Bostrom, 1971; Moore, 1973; Dickinson, 1978; Doglioni, 1994; Doglioni et al., 1999). If this is the case today, then it should have been the case throughout Earth history because of the constancy of eastward planetary rotation. Therefore, models for ancient arc-trench systems must account for the azimuth of their facing directions when they were active. Lack of recognition of this fundamental characteristic of arc-trench systems has resulted in many invalid analog models of ancient mountain belts (Dickinson, 2008).

Dickinson (1974a, 1974b), Ingersoll (1988), Ingersoll and Busby (1995) and Dorobek (2008) summarized tectonic settings and subsidence mechanisms of the diverse basin types related to arc-trench systems. Ingersoll and Busby (1995), and Smith and Landis (1995) also discussed construction and erosion of arc edifices that provide most sediment to neighboring basins.

The distinction of forearc, intra-arc, and backarc basins is not always clear. Intra-arc basins are defined as thick volcanic-volcaniclastic and other sedimentary accumulations along the arc platform, which is formed of overlapping or superposed volcanoes. The presence of vent-proximal volcanic rocks and related intrusions is critical to the recognition of intra-arc basins in the geologic record, since arc-derived volcaniclastic material may be spread into forearc, backarc, and other basins. A more general term, "arc massif," refers to crust generated by arc magmatic processes (Dickinson, 1974a, 1974b), and arc crust may underlie a much broader region than the arc platform. The distinction of forearc and intra-arc basins is also discussed by Dickinson (1995). Many backarc basins form by rifting within the arc platform (Marsaglia, 1995), and were intra-arc basins in their early stages. Also, forearc, intra-arc and backarc settings change temporally and are superposed on each other due to both gradual evolution and sudden reorganization of arc-trench systems resulting from collisional events, plate reorganization, and changes in plate kinematics.

Trenches

Karig and Sharman (1975), Schweller and Kulm (1978), Thornburg and Kulm (1987), and Underwood and Moore (1995) summarized the

dynamic nature of sedimentation and tectonics in active trenches (Fig. 1.4a). The sediment wedge of a trench is in dynamic equilibrium when subduction rate and angle, sediment thickness on the oceanic plate, rate of sedimentation, and distribution of sediment within the trench are constant. Thornburg and Kulm (1987) provided documentation of the dynamic interaction of longitudinally
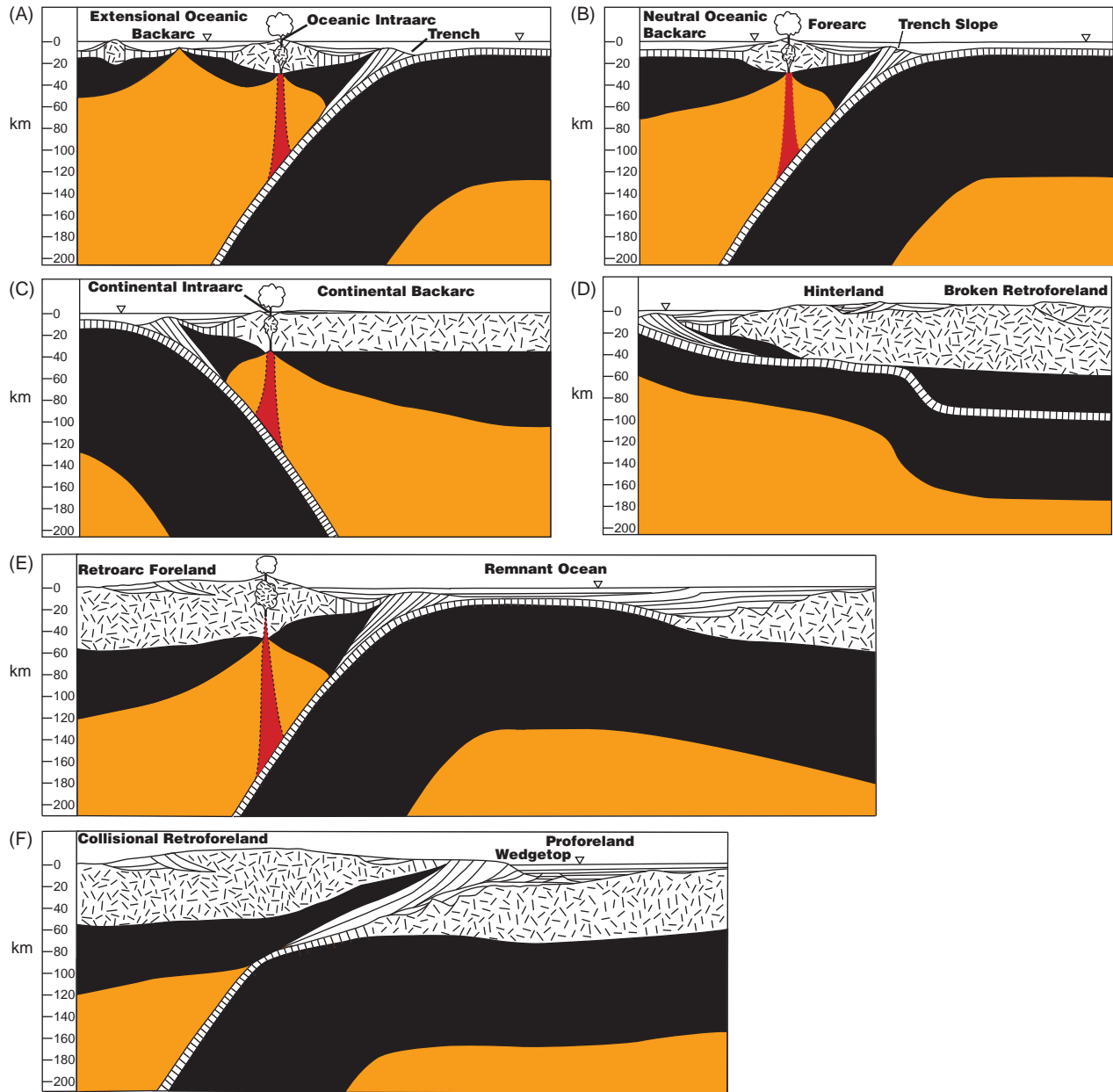
Fig. 1.4. True-scale actualistic analog models for sedimentary basins in convergent settings. A remnant arc is shown on the left side of (A). Trench, trench-slope and forearc basins are labeled only in (A and B), but they are associated with all types of arc-trench systems. Intra-arc basins may be associated with any magmatic arc, but they are more common and more likely to be preserved in extensional and neutral settings (A, B, and C). Hinterland basins may form in compressional arc-trench systems (D, E), or in collisional systems (F and Figure 1.6A–B). Remnant ocean basins form between any colliding crustal margins; a compressional arc-trench system is shown converging with an intraplate margin in (E). Wedgetop basins may form in any compressional setting; a proforeland example is shown in (F). If neutral continental arc-trench systems (C) become extensional, then they may evolve into extensional oceanic systems (A). Symbols same as in Figure 1.3; slab-generated melts = red.

transported material (trench wedge with axial channel) and transversely fed material (trench fan). With increasing transverse supply of sediment to the trench, the axial channel of the trench wedge is forced seaward and the trench wedge widens. Contrasts in dynamic trench-fill processes help determine not only trench bathymetry and depositional systems, but also accretionary architecture (Thornburg and Kulm, 1987; Underwood and Moore, 1995). This dynamic model may be useful in reconstruction of sedimentary and tectonic processes in trenches, as expressed in ancient subduction complexes.

Scholl et al. (1980) developed conceptual models relating accretionary processes to subduction and sedimentary parameters that influence forearc and trench characteristics. Cloos and Shreve (1988a, 1988b) developed quantitative models for processes at greater depths in subduction zones, which affect the nature of deformation and metamorphism, and the overall character of forearcs. Reconstruction of sedimentary systems within the transient settings of ancient trenches is highly problematic because of difficulty of studying modern systems at such great water depths, contrast in scale of resolution between modern and ancient studies, and extreme structural deformation that occurs within subduction environments (Underwood and Moore, 1995). Nonetheless, advances in technology and continuing studies of modern and ancient systems are providing incremental improvements in our understanding of the sedimentary and tectonic systems (e.g., Maldonado et al., 1994; Mountney and Westbrook, 1996; Leverenz, 2000; Kopp and Kukowski, 2003).

Trench-slope basins

Moore and Karig (1976) developed a model for sedimentation in small ponded basins along inner trench walls (Fig. 1.4b). Deformation within and on subduction complexes results in irregular bathymetry; turbidites are ponded within resulting trench-slope basins. Average width, sediment thickness, and age of basins increase up slope due to progressive uplift of deformed material and widening of fault spacing during dewatering and deformation of offscraped sediment. In ancient subduction complexes, trench-slope basins are filled with relatively undeformed, locally derived turbidites surrounded by highly deformed accreted material of variable origin. Contacts

between trench-slope basins and accreted material are both depositional and tectonic. Moore and Karig's (1976) model was developed for Nias Island near Sumatra, an area of rapid accretion of thick sediments. Their model is less useful for sediment-starved forearc areas. Allen et al. (2008), and Hall and Smyth (2008) provided additional details concerning the Nias Island area and the Andaman Islands, including some alternative interpretations. Nonetheless, Moore and Karig's (1976) general principles governing the development of sedimentary basins on the lower trench slope are fundamental to reconstructing ancient subduction complexes.

Underwood and Moore (1995), Aalto and Miller (1999), Underwood et al. (2003), and Allen et al. (2008) discussed additional examples of both modern and ancient trench-slope basins, and their significance in paleotectonic reconstructions.

Forearc basins

Dickinson and Seely (1979) and Dickinson (1995) provided a classification of arc-trench systems, similar to Dewey's (1980), and outlined plate-tectonic controls governing subduction initiation and forearc development (Fig. 1.4b). Factors controlling forearc geometry include the (1) initial setting (2) sediment thickness on subducting plate (3) rate of sediment supply to trench (4) rate of sediment supply to forearc area (5) rate and orientation of subduction, and (6) time since initiation of subduction. Arc-trench gaps tend to widen through time (Dickinson, 1973) due to prograde accretion at trenches and retrograde migration of magmatic arcs following subduction initiation. Prograde accretion is especially rapid where thick sequences of sediment are accreted. The net result of widening of the arc-trench gap is the general tendency for forearc basins to enlarge through time (e.g., Great Valley forearc basin; Ingersoll, 1979, 1982; Dickinson, 1995).

Forearc basins include the following types (Dickinson and Seely, 1979; Dickinson, 1995): (1) intramassif (transitional to intra-arc) (2) accretionary (trench-slope) (3) residual (lying on oceanic or transitional crust trapped behind the trench when subduction initiated) (4) constructed (lying across the boundary of arc massif and subduction complex), and (5) composite (combination of above settings). Residual and constructed basins tend to evolve into composite basins; commonly,

this evolutionary trend is accompanied by filling and shallowing of forearc basins.

Stern and Bloomer (1992) discussed lithospheric extension along the front edge of the overriding plate at the time of initiation of the Mariana subduction zone (Oligocene). This type of extension to form new crust is only likely within 10–20 My of initiation of an intraoceanic subduction zone within old (strong) oceanic lithosphere, where slab rollback begins as soon as subduction initiates, and the weakest part of the overriding plate is near the edge. Soon after subduction begins, forearc areas are cooled by the cold subducting oceanic lithosphere; thus, mature intraoceanic forearcs tend to be underlain by cold and strong lithosphere, and resist crustal extension (i.e., Vink et al., 1984; Steckler and tenBrink, 1986; Dickinson, 1995). Normal faults are common in shallow levels of accretionary wedges (e.g., Platt, 1986; Underwood and Moore, 1995), but crustal rifting to form new crust has not been documented in any modern forearc, and is unlikely to have occurred in any ancient forearcs (Ingersoll, 2000). In contrast, arc axes of mature intraoceanic systems tend to be the weakest parts of overriding plates, and extension is accommodated by intra-arc and backarc spreading (Marsaglia, 1995).

Several recent studies of both modern and ancient forearc basins have verified the usefulness of the general models discussed by Dickinson (1995) (e.g., Einsele et al., 1994; Van der Werff, 1996; Mountney and Westbrook, 1997; Constenius et al., 2000; Trop, 2008).

Intra-arc basins

The origin of basins within magmatic arcs (Fig. 1.4a) is, in general, poorly understood, largely due to the paucity of studies that integrate volcanology, sedimentology and basin analysis (Ingersoll, 1988; Ingersoll and Busby, 1995). A further deterrent for many sedimentologists is the fact that arcs are characterized by high heat flow with steep geothermal gradients and intense magmatism, and are commonly subjected to crustal shortening at some time in their history; therefore, experience in "seeing through" the overprints of hydrothermal alteration, metamorphism and deformation is required. Fisher and Schmincke (1984), Cas and Wright (1987), Cas and Busby-Spera (1991), Fisher and Smith (1991), and Smith and Landis (1995) provided excellent summaries of knowledge prior to 1995.

Oceanic intra-arc basins

There are at least three major types of depocenters for volcanic and sedimentary accumulations within arcs (Ingersoll and Busby, 1995; Smith and Landis, 1995). Depocenters may occur in low regions between volcanoes and along their flanks, although these have high preservation potential only below sealevel (i.e., generally in oceanic arcs). Also, depocenters with high preservation potential may form when the axis of arc volcanism shifts to a new position on an oceanic arc platform, thus creating a low region between the active chain and the abandoned chain. Smith and Landis (1995) referred to both of these types of intra-arc basin as "volcano-bounded basins." They also referred to "fault-bounded basins," which are rapidly subsiding basins where tectonic structures, rather than constructional volcanic features, account for relief along the basin margins (Ingersoll and Busby, 1995).

Additional studies of sedimentation in oceanic intra-arc settings include those of Robertson and Degnan (1994), Fackler-Adams and Busby (1998), Sowerbutts and Underhill (1998), Sowerbutts (2000), Clift et al. (2005), and Busby et al. (2006).

Continental intra-arc basins

The most important mechanisms for accumulating and preserving thick stratigraphic successions in continental arcs appear to be, in descending scale (Busby-Spera, 1988b; Busby-Spera et al., 1990): (1) plate-margin-scale extension or transtension (2) extension on a more local scale during pluton or batholith emplacement, and (3) localized subsidence of calderas during large-volume ignimbrite eruptions. Plate-margin-scale extension or transtension produces belts of continental-arc sequences that are continuous or semicontinuous over hundreds to thousands of kilometers and record high rates of subsidence over tens of millions of years. The effects of extension in the roofs of plutons or batholiths (e.g., Tobisch et al., 1986) may be difficult to distinguish from plate-margin-scale extension, but the former should operate over shorter time scales (i.e., less than a few million years), and should not, by itself, produce a low-standing arc capable of trapping sediment derived from outside the arc. Continental calderas form small (10–60 km wide) but deep (1–4 km) depocenters for ignimbrite erupted during caldera collapse, as well as for volcanic and

sedimentary strata ponded within the caldera after collapse (e.g., Riggs and Busby-Spera, 1991; Lipman, 1992; Schermer and Busby, 1994). Funk et al. (2009) provided a detailed description of the Cenozoic tectonics of the intra-arc basins of Nicaragua and El Salvador.

Backarc basins

There are two types of backarc basins: (1) oceanic basins behind intraoceanic magmatic arcs, and (2) continental basins behind continental-margin arcs that lack foreland foldthrust belts (Ingersoll and Busby, 1995). Many backarc basins are extensional in origin, forming by rifting and seafloor spreading (Fig. 1.4a) (Marsaglia, 1995). These commonly originate through rifting of the arc, either along its axis (intra-arc) or immediately to the front or rear of its axis. The term "interarc basin" (Karig, 1970) has been widely superseded by the term "backarc basin," but it can be used where rifting has occurred along or near an arc axis, thus eventually producing a remnant arc behind the backarc basin. The presence or preservation of a remnant arc is not a necessary condition for recognition of a backarc basin (Taylor and Karner, 1983).

Many backarc basins are nonextensional (Marsaglia, 1995), forming under neutral stress regimes (Fig. 1.4b). The most common type of nonextensional backarc basin consists of old ocean basins trapped during plate reorganization (e.g., the Bering Sea). Also, nonextensional backarc basins develop on continental crust (Fig. 1.4c) (e.g., Sunda shelf of Indonesia). Backarc shortening may occur in intraoceanic arc-trench systems involved in early stages of collision with buoyant crust (e.g., Greater Antilles and eastern Indonesia) (tenBrink et al., 2009); this shortening could represent early stages of subduction initiation during polarity reversal.

Oceanic backarc basins

Modern oceanic backarc basins may be distinguished from other ocean basins petrologically or by their positions behind active or inactive arc-trench systems (Taylor and Karner, 1983; Marsaglia, 1995). Such diagnostic features are commonly not preserved in ancient backarc basins, which commonly undergo metamorphic and structural modifications during emplacement in orogenic zones as ophiolites. The nature and

timing of deposition of sediment on top of ophiolite sections have proven more diagnostic for determining original plate-tectonic settings (e.g., Tanner and Rex, 1979; Hopson et al., 1981, 2008; Kimbrough, 1984; Busby-Spera, 1988a; Robertson, 1989).

The most detailed study of a backarc volcaniclastic apron and its substrate comes from Middle Jurassic rocks in Mexico (Busby-Spera, 1987, 1988a; Critelli et al., 2002). That study supported Karig and Moore's (1975) assertion that oceanic backarc basins isolated from terrigenous sediment influx may show the following simple, uniform sedimentation patterns: (1) lateral and vertical differentiation of facies due to progradation of a thick volcaniclastic apron into a widening backarc basin; such an apron may extend for more than 100 km from a volcanic island and grow to a thickness of 5 km in 5 My (Lonsdale, 1975). This phase is followed by (2) blanketing of the apron with a thin sheet of mud and sand eroded from the arc after volcanism and spreading have ceased. This cycle reflects the temporal episodicity of seafloor spreading in oceanic backarc basins, which appear to form in 10–15 My or less (Taylor and Karner, 1983). As a result, extensional oceanic backarc basins generally have shorter life spans than intra-arc basins (Fig. 1.2). The shorter life span reflects temporal episodicity of extensional oceanic backarc basins, the most common type of backarc basin. In contrast, arcs may undergo episodic extension for many tens of millions of years, particularly in continental settings. Although backarc basins and their fill make an important contribution to orogenic belts, most ancient oceanic backarc basins have probably been subducted; the frontal-arc sides of backarc basins may be preferentially preserved in the geologic record (Busby-Spera, 1988a).

Marsaglia (1995) discussed modern and ancient backarc basins, both oceanic and continental, and both extensional and neutral settings, although her emphasis was on extensional backarcs of the western Pacific (also, see Klein, 1985). More recent publications addressed evolution of the complex extensional backarc basins of the Western Mediterranean (e.g., Maillard and Mauffret, 1999; Pascucci et al., 1999; Mattei et al., 2002; Rollet et al., 2002). Sibuet et al. (1998) synthesized the tectonic and magmatic evolution of the Okinawa Trough, and Critelli et al. (2002) analyzed the Jurassic backarc basin of Cedros Island, Baja California using sandstone petrology in conjunction with stratigraphy and sedimentology. Less

attention has been paid to nonextensional backarc basins (e.g., Aleutian Basin of the Bering Sea), formed by trapping of oceanic crust behind intraoceanic arcs following initiation of intraoceanic subduction zones (Ben-Avraham and Uyeda, 1983; Tamaki and Honza, 1991). Part of the Caribbean Sea, the West Philippine Basin, and perhaps part of the Okhotsk Sea north of the Kuril Basin also consist of oceanic crust trapped in backarc settings (Uyeda and Ben-Avraham, 1972; Scholl et al., 1975: Ben-Avraham and Uyeda, 1983; Marsaglia, 1995). These backarc basins generally have longer life spans and greater preservation potential than indicated in Figure 1.2, especially if they evolve into dormant ocean basins. Oceanic crust trapped in backarc settings is likely to be as complex as any other oceanic crust, with oceanic plateaus, continental fragments, and transform faults (Marsaglia, 1995).

Continental backarc basins

A modern continental backarc in a neutral stress regime is the Sunda Shelf of Indonesia (Hamilton, 1979; Ingersoll, 1988; Ingersoll and Busby, 1995). DeCelles and Giles (1996) utilized the Sunda Shelf as an example of the earliest stages of development of a retroforeland basin (see below), but Moss and McCarthy (1997) disputed this interpretation and suggested that there is no retroarc shortening behind the Indonesian magmatic arc (also see DeCelles and Giles, 1997). Moss and McCarthy (1997) interpreted part of the Sunda Shelf area as having a previous extensional history. In any case, an extensional backarc can evolve into a neutral backarc, which can evolve into a retroforeland basin. Variable stress regimes in backarc and retroarc settings are common.

A similar series of backarc-to-retroforeland basins developed during the Mesozoic in the western USA (Dickinson, 1981a, 1981b; Lawton, 1994; Ingersoll, 1997, 2008a). Following the Permian-Triassic Sonoma orogeny, a continental-margin magmatic arc developed following subduction initiation (Hamilton, 1969; Schweickert, 1976, 1978; Busby-Spera, 1988b). Shallow-marine and non-marine conditions prevailed in the dynamically neutral backarc area from mid-Triassic to Late Jurassic (Dickinson, 1981a, 1981b; Lawton, 1994; Ingersoll, 1997, 2008a), although crustal extension may have characterized some parts of the backarc (Wyld, 2000, 2002). The backarc evolved into a retroforeland as shortening initiated during the

Jurassic (Oldow, 1984; Oldow et al., 1989; Lawton, 1994; Wyld, 2002). Thus, relative timing of extension, neutrality, and shortening in the Sunda Shelf area is similar to the interpreted sequence of events in the Mesozoic backarc-retroforeland of the western USA.

Retroforeland basins

Compressional arc-trench systems commonly develop foreland basins behind arcs due to partial subduction of continental crust beneath arc orogens (Dickinson, 1974b; Dewey, 1980; Ingersoll, 1988; Ingersoll and Busby, 1995; DeCelles and Giles, 1996). "Foreland basin" is a pre-plate-tectonic term used to describe a basin between an orogenic belt and a craton (Allen et al., 1986). Dickinson (1974b) proposed that the term "retroarc" be used to describe foreland basins formed behind compressional arcs, in contrast to "peripheral" foreland basins formed on subducting plates during continental collisions. Thus, although "backarc" and "retroarc" are literally synonymous, the former is used for extensional and neutral arc-trench systems, whereas the latter is used for compressional arc-trench systems.

Willett et al. (1993), Johnson and Beaumont (1995), Beaumont et al. (1996), and Naylor and Sinclair (2008) modified Dickinson's (1974b) original nomenclature for foreland basins by shortening "retroarc foreland" to "retroforeland" and changing "peripheral foreland" to "proforeland". This nomenclature is adopted herein, with the clear stipulation that retroforeland basins form on the upper plates of convergent margins and proforeland basins form on the lower plates of convergent margins. Retroforelands tend to have longer histories than proforelands because the former commonly initiate during subduction of oceanic lithosphere (e.g., Andean retroforeland), whereas proforelands do not exist until buoyant continental crust enters subduction zones (inducing collision) (Dickinson, 1974b; Ingersoll, 1988; Cloos, 1993; Ingersoll and Busby, 1995). In order to clarify these distinctions, I propose that retroforelands be subdivided into retroarc forelands (formed behind continental-margin arcs, e.g., the Andes) and collisional retroforelands (formed on the overriding continental plate during continental collision, e.g., South Alpine foreland basin). The general term "retroforeland" may be used for any foreland on the upper plate of a convergent margin, whereas the more restricted terms would

be used to designate whether oceanic or continental crust was being subducted on the opposite side of the orogen. Approximately one third of active magmatic arcs have associated retroarc forelands, whereas arc activity commonly ceases as profore-lands and collisional retroforelands develop.

Retroarc foreland basins

Jordan (1981) presented an analysis of the asymmetric Cretaceous retroarc foreland basin associated with the Idaho-Wyoming thrust belt. She used a two-dimensional elastic model to show how thrust loading and sedimentary loading resulted in broad flexure of the lithosphere (Fig. 1.4e). The location of maximum flexure migrated eastward as thrusting migrated eastward. The area of subsidence broadened due to erosional and depositional redistribution of part of the thrust load, and possibly enhanced by high eustatic sea level of the Late Cretaceous. Comparison of modeled basin and basement geometries with isopach maps provides tests of possible values of flexural rigidity of the lithosphere. The modern sub-Andean thrust belt and foreland basin have similar topography to that proposed for the Cretaceous of the Idaho-Wyoming system (Jordan, 1995). Topography is controlled by thrust-fault geometry and isostatic subsidence.

The models presented by Jordan (1981) and Beaumont (1981) are broadly applicable to other retroarc foreland basins (Jordan, 1995; DeCelles and Giles, 1996; Catuneanu, 2004). These and derivative models demonstrate that tectonic activity in foreland foldthrust belts is the primary cause of subsidence in associated foreland basins (Price, 1973). Sedimentary redistribution, autocyclic sedimentary processes, dynamic effects of asthenospheric circulation (e.g., Gurnis, 1993; Burgess et al., 1997), and eustatic sealevel changes are important modifying factors in terms of regressive-transgressive sequences, but compressional tectonics behind the arc-trench system is the driving force. The Cretaceous seaway of North America was largely the result of this compressional tectonic activity (combined with high eustatic sealevel) (Dickinson, 1976b, 1981a). Details concerning timing of thrusting and initial sedimentary response to thrusting within the Idaho-Wyoming thrust belt have been debated (e.g., Heller et al., 1986), but the essential role of compressional tectonics in creating retroarc foreland basins is clear (Price, 1973; DeCelles and Giles, 1996).

Jordan (1995) updated her analysis of the Cretaceous retroarc foreland of North America, synthesized the Neogene to Holocene retroarc foreland of South America, and discussed general models for retroarc foreland basins. DeCelles and Giles (1996) synthesized foreland-basin systems, including subdivision into four discrete depozones: wedge-top, foredeep, forebulge, and backbulge. All four depozones occur in the modern retroforeland east of the central Andes (Horton and DeCelles, 1997). DeCelles and Horton (2003) applied this model to interpretation of tertiary foreland strata of Bolivia, and concluded that approximately 1000 km of foreland crust have been thrust westward beneath the Andean orogenic belt. Fildani et al. (2003), Abascal (2005), Gomez et al. (2005), Hermoza et al. (2005), Horton (2005), and Uba et al. (2005) provided detailed analyses of diverse parts of the Andean retroforeland system.

DeCelles and Giles's (1996) subdivisions apply equally to all types of foreland-basin systems. Wedgetop basins are discussed separately below because their character is tied directly to fault dynamics, whereas the other three depozones are created by flexural loading of the overall thrust belt. Wedgetop and backbulge depozones have not been considered in most models for foreland evolution (DeCelles and Giles, 1996). Foreland models that include wedgetop depozones must utilize doubly tapered prisms in cross section, rather than the wedge that is commonly used as a simplification (DeCelles and Giles, 1996). Dorobek and Ross (1995) illustrated many types of models and case studies that improve our understanding of foreland basins.

Collisional retroforeland basins

The South Alpine collisional retroforeland basin developed synchronously with Alpine orogenesis as the European plate subducted beneath Adria (Bertotti et al., 1998; Carrapa, 2009) (Fig. 1.4f). Late Cretaceous evolution of this foreland began in a retroarc setting, but developed into a collisional retroforeland concurrent with Alpine orogenesis and development of the better known North Alpine (molasse) proforeland basin (Bertotti et al., 1998). Bertotti et al. (1998) suggested progressive weakening of the flexed Adria plate through time. Carrapa and Garcia-Castellanos (2005) demonstrated that the Tertiary Piedmont basin of the western Po Plain formed by Alpine retroforeland flexure of a visco-elastic plate during

Oligocene-Miocene time. Zattin et al. (2003) used provenance data from the Venetian basin to document the sequence of deformation in the eastern South Alpine collisional retroforeland. Apennine orogenesis has superposed the Po Valley proforeland basin on the older retroforeland basin, thus making this a hybrid foreland basin (Ingersoll and Busby, 1995; Miall, 1995).

The Triassic-Jurassic foreland sequences of the Ordos basin of central China represent deposition in a collisional retroforeland basin related to suturing of the North and South China blocks (Sitian et al., 1995; Ritts et al., 2009).

Broken retroforeland basins

Low-angle subduction beneath compressional arc-trench systems may result in basement-involved deformation within retroarc foreland basins (Fig. 1.4d) (Dickinson and Snyder, 1978; Jordan, 1995). The Rocky Mountain region of the western USA is the best-known ancient example of this style of deformation; similar modern provinces have been documented in the Andean foreland (e.g., Jordan et al., 1983a, 1983b; Jordan and Allmendinger, 1986; Jordan, 1995).

Chapin and Cather (1981), Dickinson et al. (1988, 1990), Cather and Chapin (1990), Dickinson (1990), Hansen (1990), and Lawton (2008) synthesized and discussed controls on latest Cretaceous through Eocene (Laramide) sedimentation and basin formation of the Colorado Plateau and Rocky Mountain area. They agreed that diverse types of uplifts and basins formed during this period, but they disagreed on paleodrainage networks, the relative importance of strike-slip deformation along the east side of the Colorado Plateau, and whether the Laramide orogeny occurred in two distinct stages or was a continuum of responses to a generally homogeneous strain field. Yin and Ingersoll (1997) and Ingersoll (2001) presented a model for Laramide crustal strain and basin evolution in northern New Mexico and southern Colorado, which is consistent with a generally homogeneous strain field. Hoy and Ridgway (1997) illustrated the complex structural, stratigraphic and sedimentologic relations that commonly developed along the margins of intraforeland uplifts in Wyoming. Cardozo and Jordan (2001), Davila and Astini (2003), Sobel and Strecker (2003), and Hilley and Strecker (2005) studied broken retroforeland basins and associated uplifts in Argentina.

Remnant ocean basins

Intense deformation occurs in suture belts during the attempted subduction of buoyant (nonsubductable) continental or magmatic-arc crust (e.g., Cloos, 1993). Suture belts can involve rifted continental margins and continental-margin magmatic arcs (terminal closing of an ocean basin) or various combinations of arcs and continental margins (Fig. 1.4e–f). Colliding continents tend to be irregular, and great variability of timing, structural deformation, sediment dispersal patterns and preservability occurs along strike (Dewey and Burke, 1974).

Graham et al. (1975) and Ingersoll et al. (1995, 2003) used Cenozoic development of the Himalayan-Bengal system as an analog for late Paleozoic development of the Appalachian-Ouachita system, and proposed a general model for sediment dispersal related to sequentially suturing orogenic belts. "Most sediment shed from orogenic highlands formed by continental collisions pours longitudinally through deltaic complexes into remnant ocean basins as turbidites that are subsequently deformed and incorporated into the orogenic belts as collision sutures lengthen" (Graham et al., 1975, 273). This model provides a general explanation for many synorogenic flysch and molasse deposits associated with suture belts, although many units called "flysch" and "molasse" have different tectonic settings (Ingersoll et al., 1995, 2003; Miall, 1995).

North American examples of arc-continent collisions, with variable volumes of remnant-ocean-basin flysch, include the Ordovician Taconic orogeny of the Appalachians (e.g., Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985; Lash, 1988; Bradley, 1989; Bradley and Kidd, 1991) and the Devonian-Mississippian Antler orogeny of the Cordillera (e.g., Speed and Sleep, 1982; Dickinson et al., 1983). In both cases, it is difficult to clearly distinguish remnant ocean basins from incipient proforeland basins as the depositional sites for "flysch" (Ingersoll and Busby, 1995; Ingersoll et al., 1995; Miall, 1995).

Ingersoll et al. (1995, 2003) reviewed several remnant ocean basins, both modern and ancient, and demonstrated that submarine fans in remnant ocean basins represent the largest accumulations of sediment on Earth. The modern Bengal Fan is the largest sediment body and the Indus Fan is second largest; both are derived from the greatest uplifted area on Earth, the Tibetan Plateau and

Himalaya. The Triassic Songpan-Ganzi complex of northern Tibet and the Carboniferous-Permian Ouachita-Marathon flysch of Arkansas, Oklahoma, and Texas also were deposited in remnant ocean basins adjoining uplifted continental suture belts, and they are of comparable size to the Bengal and Indus fans, although deformation during suturing makes their reconstruction difficult (Ingersoll et al., 2003). No known or suggested mechanism can produce sediment masses of comparable volume. Many additional examples of remnant ocean basins, associated with both continent-continent and continent-intraoceanic-arc collisions, are discussed by Ingersoll et al. (2003).

Proforeland basins

As continental collision occurs between a rifted continental margin and the subduction zone of an arc-trench system, a tectonic load is placed on the rifted margin, first below sealevel, and later subaerially (Dickinson, 1974b; Ingersoll, 1988; Miall, 1995). A proforeland basin forms as the elastic lithosphere flexes under the encroaching dynamic load (Fig. 1.4f). Normal faulting in front of the dynamic load and uplift of a forebulge are initial responses to flexure as the dynamic load encroaches on the foreland (Bradley and Kidd, 1991; Miall, 1995; DeCelles and Giles, 1996).

Discrimination of ancient proforelands and collisional retroforelands (Fig. 1.4f) is difficult, but may be possible based on the following characteristics (Ingersoll, 1988; Ingersoll and Busby, 1995): (1) polarity of magmatic arc (2) presence of oceanic subduction complex associated with earliest phases of proforeland (3) greater water depths in proforeland (foredeep stage) (4) asymmetry of suture belt (closer to proforeland) (5) protracted development of retroforeland (longterm arc evolution) versus discrete development of proforeland (terminal ocean closure without precursor), and (6) possible volcaniclastic input to retroforeland, especially during early development, versus minimal volcaniclastic input to proforeland.

Stockmal et al. (1986) provided a dynamic 2D model for the development of proforeland basins, following finite times of rifting. They modified the model of Speed and Sleep (1982), and demonstrated the effects of rifted-margin age and topography on lithospheric flexure and basin development. The primary effect of age shows up as a higher flexural forebulge and thicker trench fill during earlier stages of attempted

subduction of an old (120 my) margin. Subsequent development is relatively insensitive to margin age. Foreland-basin subsidence is sensitive to overthrust load, with depths possibly exceeding 10 km. Crustal thickness may reach 70 km during the compressional phase (e.g., Himalayas). Tens of kilometers of uplift and erosion, of both the allochthon and the proximal foreland basin, are predicted during and after deformation. Most erosional detritus is deposited elsewhere due to uplift within the foreland; longitudinal transport into remnant ocean basins results (Graham et al., 1975; Ingersoll et al., 1995, 2003; Miall, 1995). Thick overthrusts with low topographic expression are to be expected where broad, attenuated rifted continental margins have been pulled into subduction zones (Stockmal et al., 1986).

Miall (1995) discussed all "collision-related foreland basins," which include both proforeland and collisional retroforeland basins. Several studies of foreland basins were presented in Dorobek and Ross (1995) and Mascle et al. (1998). The rapidly expanding literature on proforeland basins includes study of the Pyrenean (e.g., Arenas et al., 2001; Jones et al., 2004), Alpine (e.g., Sinclair, 1997; Gupta and Allen, 2000; Allen et al., 2001; Pfiffner et al., 2002; Kempf and Pfiffner, 2004), Apennine (e.g., Bertotti et al., 2001; Lucente, 2004), Carpathian (e.g., Zoetemeijer et al., 1999; Tarapoanca et al., 2004; Leever et al., 2006), Zagros (e.g., Alavi, 2004), Himalayan (e.g., Pivnik and Wells, 1996; DeCelles et al., 1998, 2001; Najman and Garzanti, 2000; Najman et al., 2004), Longmen Shan (e.g., Yong et al., 2003; Meng et al., 2005), West Taiwan (e.g., Chen et al., 2001; Lin and Watts, 2002), Papuan (e.g., Galewsky et al., 1996; Haddad and Watts, 1999), Appalachian (e.g., Thomas, 1995; Castle, 2001), and Proterozoic (e.g., Saylor, 2003) forelands.

Wedgetop basins

Ori and Friend (1984) defined "piggyback basins" as basins that form and fill while being carried on moving thrust sheets. DeCelles and Giles (1996) suggested "wedgetop" as a more general descriptive term, which includes both "piggyback" and "thrust-top" basins (Fig. 1.4f). Wedgetop basins are dynamic settings for sediment accumulation; most sediment is derived from associated foldthrust belts, with subordinate input from arc and basement terranes (Critelli and Le Pera, 1994; Trop and

Ridgway, 1997). The foldthrust belts can be in proforeland, retroforeland, or transpressional settings (Ingersoll and Busby, 1995). Wedgetop basins share characteristics with trench-slope basins. The submarine environment of southern Taiwan illustrates the transition from forearc/trench-slope/trench environments west of the Luzon Arc to orogenic-wedge/wedgetop/foredeep of the Taiwan collision zone (Chiang et al., 2004). This transition occurs where subduction of oceanic crust beneath the Luzon Arc evolves into attempted subduction of Asian continental crust to form the Taiwan suture zone. Both trench-slope and wedgetop basins have low preservation potential due to their development on growing thrust belts; therefore, they are generally found only in young orogenic systems (e.g., Burbank and Tahirkheli, 1985) (Fig. 1.2).

Jordan (1995), Miall (1995), Nilsen and Sylvester (1995), and Talling et al. (1995) provided additional insights regarding wedgetop basins and their relations with compressional processes. Foreland deposition can evolve into wedgetop deposition as frontal faults prograde into foreland basins (DeCelles and Giles, 1996; Pivnik and Khan, 1996); where rapid "sled run" advance of the thrust front occurs due to low basal friction, most foreland deposition may occur in diverse wedgetop basins (Evans and Elliott, 1999; Ford, 2004). Horton (1998) documented extensive Oligocene to Upper Miocene synorogenic wedgetop deposits in the Eastern Cordillera of southern Bolivia. He suggested that longterm preservation of these deposits (for up to 30 My) may reflect the semiarid climate of the region; their presence adds mass to the orogenic wedge, which, in turn, promotes propagation of the thrust front (Horton, 1998). Wedgetop basins form in both thin-skinned and thick-skinned (basement-involved) contractional settings (e.g., Casas-Sainz et al., 2000).

Hinterland basins

Horton (chapter 21, this volume) described two classes of hinterland basins: those formed in non-collisional retroarc orogens (e.g., Andes) and collisional orogens (e.g., Himalaya-Tibet). Because "hinterland" refers to the "internal" parts of orogens, opposite the direction of vergence of folds and faults, the term denotes a direction relative to a foldthrust belt. In a literal sense, all parts of an orogen behind a foldthrust belt constitute the hinterland, regardless of genetic origin; for example, a retroforeland is part of the hinterland of a proforeland, and vice versa, in two-sided orogens (e.g., Alps and Pyrenees). Horton (chapter 21, this volume) and I define "hinterland basin" in the more restricted sense of basins within orogenic belts that do not fall into any other category described herein.

Hinterland basins record nonmarine sedimentation, usually at high elevations, that formed on thickened continental crust (Horton, chapter 21, this volume) (Fig. 1.4d). As a result, they have low preservation potential, and relatively short life spans (Fig. 1.2), although some basins have life spans of tens of millions of years (e.g., Altiplano; Horton et al., 2002). Extensional, contractional, and strike-slip processes can create the accommodation space for hinterland basins, with fault-induced crustal thinning, sedimentary and volcanic loading, and tectonic loading causing subsidence (Fig. 1.1). Horton (chapter 21, this volume) described two modes of hinterland-basin evolution: (1) basins that developed as new faults became active, and (2) basins that overprinted former foreland basins as the deformation front advanced.

Additional examples of hinterland basins are discussed by Burchfiel et al. (1992), Garzione et al. (2003), Alcicek (2007), DeCelles et al. (2007), Giovanni et al. (2010), and Saylor et al. (2010).

TRANSFORM SETTINGS

Strike-slip systems

The complexity and variability of sedimentary basins associated with strike-slip faults are almost as great as for all other types of basins (Ingersoll and Busby, 1995). Transform faults in oceanic lithosphere generally behave according to the plate-tectonic model, whereas strike-slip faults in continental lithosphere are extremely complex and difficult to fit into a model involving rigid plates.

Strike-slip faults within continental crust are likely to experience alternating periods of extension and compression as slip directions adjust along major crustal faults (Crowell, 1974a, 1974b; Reading, 1980). Thus, opening and closing of basins along strike-slip faults is analogous, at smaller spatial and time scales, to the opening and closing of ocean basins (the Wilson Cycle) (e.g., Wilson, 1966; Dewey and Burke, 1974). This process is illustrated beautifully by the Neogene to Holocene development of southern

California (e.g., Crowell, 1974a, 1974b; Schneider et al., 1996; Ingersoll and Rumelhart, 1999; Kellogg and Minor, 2005; Ingersoll, 2008b).

Basins related to strike-slip faults can be classified into end-member types, although most basins are hybrids. Transtensional (including pull-apart) basins form near releasing bends and transpressional basins form at constraining bends (Crowell, 1974b). Basins associated with crustal rotations about vertical axes within the rotating blocks ("transrotational"; Ingersoll, 1988) may experience any combination of extension, compression, and strike slip (Ingersoll and Busby, 1995).

Christie-Blick and Biddle (1985) and Nilsen and Sylvester (1995) reviewed structural and stratigraphic development of strike-slip basins, based, in large part, on the pioneering work of Crowell (1974a, 1974b). They illustrated structural complexity along strike-slip faults, and implications for associated basins. Primary controls on structural patterns are the (1) degree of convergence and divergence of adjacent blocks (2) magnitude of displacement (3) material properties of deformed rocks, and (4) preexisting structures (Christie-Blick and Biddle, 1985). Subsidence in sedimentary basins results from crustal attenuation, thermal subsidence during and following extension, flexural loading due to compression, and sedimentary loading. Thermal subsidence is faster, but less in total magnitude in narrow transtensional basins than in elongate orthogonal rifts due to lateral heat conduction in the former. Distinctive aspects of sedimentary basins associated with strike-slip faults include (Christie-Blick and Biddle, 1985) (1) mismatches across basin margins (2) longitudinal and lateral basin asymmetry (3) episodic rapid subsidence (4) abrupt lateral facies changes and local unconformities, and (5) marked contrasts in stratigraphy, facies geometry, and unconformities among different basins in the same region.

These characteristics of strike-slip systems have been documented by studies of both modern and ancient, and onshore and offshore fault systems (e.g., Barnes et al., 2001, 2005; Koukouvelas and Aydin, 2002; Hsiao et al., 2004; Okay et al., 2004; Seeber et al., 2004; Wakabayashi et al., 2004).

Transtensional basins

Transtensional basins (Fig. 1.5a) form at left-stepping sinistral fault junctures and at right-stepping dextral fault junctures (Crowell, 1974a, 1974b; Reading 1980; Christie-Blick and Biddle,

1985; Nilsen and Sylvester, 1995). Mann et al. (1983) proposed a model for such basins based on a comparative study of pull-apart basins at various stages of development. Pull-apart basins evolve through the following stages: (1) nucleation of extensional faulting at releasing bends of master faults; (2) formation of spindle-shaped basins defined and commonly bisected by oblique-slip faults; (3) further extension, producing "lazy-S" or "lazy-Z" basins; (4) development into rhombo-chasms, commonly with two or more sub-circular deeps; and (5) continued extension, resulting in the formation of oceanic crust at short spreading centers offset by long transforms. Basaltic volcanism and intrusion may be important during stages 3 through 5 (e.g., Crowell, 1974b). Most pull-apart basins have low length-to-width ratios, due to their short histories in changing strike-slip regimes (Mann et al., 1983). Mann (1997) demonstrated how large transtensional basins commonly form in zones of tectonic escape. Long-lived transtensional plate margins may evolve into transtensional nascent ocean basins (e.g., Gulf of California) or intraplate transform continental margins (e.g., south coast of West Africa).

Physical analog modeling provides important insights concerning initiation and evolution of pull-apart basins (e.g., Dooley and McClay, 1997; Rahe et al., 1998). Integrated geophysical and geologic studies have been conducted on several young transtensional basins and fault zones in both subaerial and submarine environments, for example: Hope fault of New Zealand (Wood et al., 1994), Dead Sea transform (Katzman et al., 1995; Hurwitz et al., 2002; Lazar et al., 2006), and North Anatolian fault in Sea of Marmara (Okay et al., 1999; Rangin et al., 2004). Dorsey et al. (1995) discussed the effects of rapid fault-controlled subsidence on fan-delta sedimentation along the margin of the transtensional Gulf of California. Waldron (2004) demonstrated that the Middle Pennsylvanian Stellarton pull-apart basin of Nova Scotia had a complex history of multiple overprinted structures. The overall structural and stratigraphic development of this ancient transtensional feature is consistent with the models of Mann et al. (1983), Dooley and McClay (1997), and Rahe et al. (1998).

Transpressional basins

Transpressional basins (Fig. 1.5b) include two types: 1. severely deformed and overthrust margins along sharp restraining bends that result in flexural
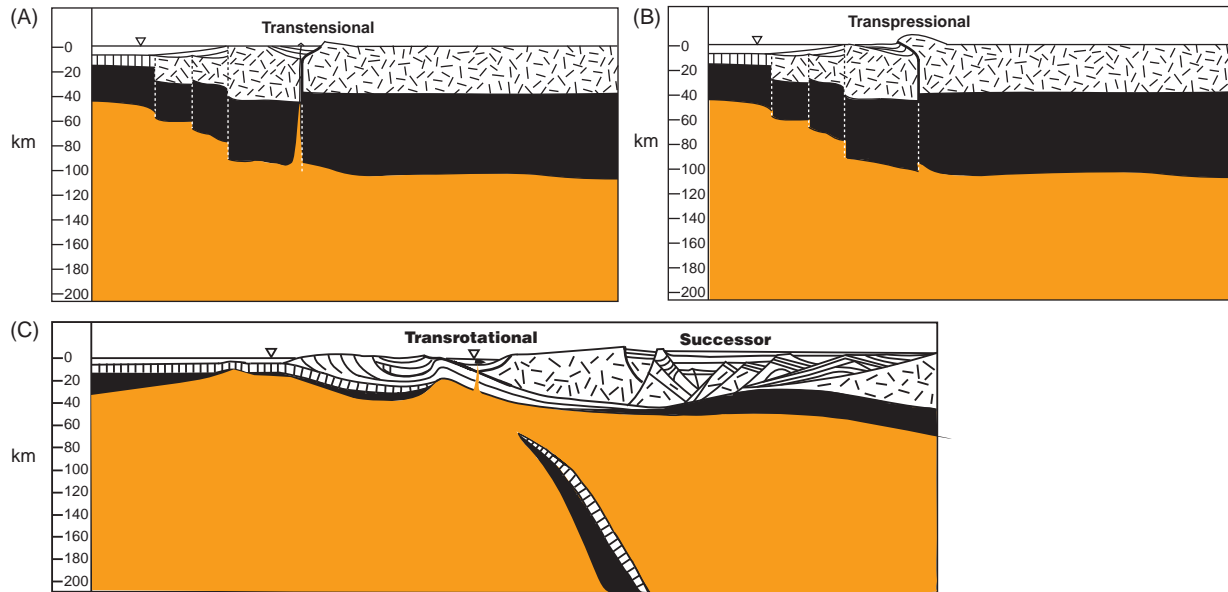
Fig. 1.5. True-scale actualistic analog models for sedimentary basins in transform and miscellaneous settings. These cross sections are based on the Miocene to Holocene evolution of southern California (Ingersoll, 2008b); many other tectonic settings are common for transtensional and transpressional basins. Transrotational basins are less common; the three-plate interactions depicted in (C) are unique on Earth today. Successor basins may form following cessation of any tectonic activity; shown in (C) is the southern Basin and Range of the USA, where undeformed Neogene to Holocene strata overlie tilted fault blocks and strata of the older extensional regime. Symbols same as in Figure 1.3.

subsidence due to tectonic load (e.g., northern Los Angeles basin, southern California; Schneider et al., 1996); and 2. fault-wedge basins at gentle restraining bends that result in rapid uplift of one or two margins and rapid subsidence of a basin as one block moves past the restraining bend (e.g., Neogene Ridge basin, southern California) (Crowell, 1974b, 2003a, 2003b). A basin model for type 1 would involve flexural loading similar to the foreland models discussed above, although at smaller scale.

Ridge basin is one of the most elegantly exposed and carefully studied transpressional basins in the world, as summarized by Crowell and Link (1982) and Crowell (2003a). Crowell (2003b) presented a dynamic model for the evolution of Ridge basin (12–5 Ma), a narrow crustal sliver caught between the San Gabriel fault to the southwest, and northwest-trending faults that became active sequentially in a northeast direction on the northeast side of the basin. Ridge basin became inactive when motion was transferred completely to the modern San Andreas fault (Crowell and Link, 1982; Ingersoll and Rumelhart, 1999; Crowell, 2003b). As a result of movement on the San Gabriel fault, the southwest side of the basin was uplifted and the Violin Breccia was deposited

along the basin margin. The depressed floor of the basin moved past this uplifted margin, while concurrently receiving abundant sediment from the northeast. Older depocenters moved southeastward past the restraining bend, after receiving sediment in conveyor-belt fashion, with uplift and tilting following deposition. The result is a stratigraphic thickness of over 11 km in outcrop, although vertical thickness of the basin fill is approximately one third of this. Many extraordinarily thick coarse clastic units in ancient, narrow fault-bounded basins likely were deposited in similar settings. (See Ingersoll and Busby, 1995, for a discussion of May et al.'s [1993] rejected transtensional model for the development of Ridge basin.)

McClay and Bonora (2001) developed analog models for restraining stepovers. Several studies of young transpressional features have been completed in both submarine and subaerial settings, for example the Alpine fault, New Zealand (Norris and Cooper, 1995; Barnes et al., 2005); Kobe and northern Osaka basins, Japan (Itoh et al., 2000), Cibao basin, Hispaniola (Erikson et al., 1998); Maturin foreland basin, Venezuala (Jacome et al., 2003); and southern Falkland basin (Bry et al., 2004). Trop et al. (2004) documented a transpressional origin for the Oligocene Colorado Creek basin

along the Denali fault system of Alaska. Meng et al. (2005) demonstrated how the northwest Sichuan basin (south China) evolved from a pro-foreland basin into a transpressional basin during the Mesozoic.

Transrotational basins

Paleomagnetic data from southern California document extensive clockwise rotation of several crustal blocks (more than 90 degrees locally), beginning in the Miocene and continuing today (e.g., Luyendyk et al., 1980; Hornafius et al., 1986; Luyendyk, 1991). Luyendyk and Hornafius (1987) developed their geometric model in order to make testable predictions concerning amount and direction of slip on faults bounding rotated and non-rotated blocks, and areas of gaps (basins) and overlap (overthrusts) among blocks. Dickinson (1996) quantified the amount of cumulative slip along the San Andreas transform fault system that is contributed by transrotational tectonism in southern California. Recognition of this contribution helps resolve discrepancies between Pacific-North American plate motions, and demonstrated offset along and within the North American continental margin (Dickinson and Wernicke, 1997).

Nicholson et al. (1994) developed a model of microplate capture that explains how complex interactions among the North American, Pacific and Farallon plates, starting soon after 30 Ma (Atwater, 1970, 1989; Bohannon and Parsons, 1995) led to three distinct phases of transfer of sections of coastal southern California onto the Pacific plate. The first phase (18–12 Ma) resulted in rapid clockwise vertical-axis rotation, with accumulation of the Topanga Formation in complex supradetachment basins (Ingersoll and Rumelhart, 1999; Ingersoll, 2008b) (Fig. 1.5c). Crouch and Suppe (1993) proposed that large-magnitude, core-complex-style extension formed in the wake of the rotating western Transverse Ranges. The southern California borderland and Los Angeles basin are floored by the Catalina Schist, interpreted by Crouch and Suppe (1993) as a footwall metamorphic tectonite, tectonically denuded below a detachment.

A model that successfully explains the extraordinarily complex basins of the Los Angeles area will need to integrate the transrotational models of Luyendyk and Hornafius (1987) and Dickinson (1996), the detachment model of Crouch and Suppe (1993), the microplate-capture model of Nicholson et al. (1994), and the detailed stratigraphic, sedimentologic and structural history of the Los Angeles and related basins (e.g., Wright, 1991; Ingersoll and Rumelhart, 1999; Ingersoll, 2008b).

MISCELLANEOUS AND HYBRID SETTINGS

Aulacogens

During continental rifting, three rifts commonly form at approximately 120 degrees, probably because this is a least-work configuration (Burke and Dewey, 1973). Regardless of whether initiating processes are "active" or "passive" (i.e., Sengor and Burke, 1978; Morgan and Baker, 1983), in the majority of cases, two rift arms proceed through the stages of continental separation, whereas seafloor spreading fails to develop in the third aim, resulting in a fossil rift (Sengor, 1995). Hoffman et al. (1974) discussed resulting sedimentary basins, with emphasis on a Proterozoic example. They outlined five developmental stages of the Athapuscow aulacogen, which with slight modification, provide a model applicable to most aulacogens (linear sedimentary troughs at high angles to orogens) (Fig. 1.6a): the (1) rift stage (2) transitional stage (3) downwarping stage (4) reactivation stage, and (5) postorogenic stage.

Sengor (1995) demonstrated the diverse ways in which fossil rifts (precursors to aulacogens) form, including doming, rifting, and drifting (Hoffman et al., 1974), membrane stresses, rift-tip abandonment, and continental rotation. All of these processes can result in "a narrow, elongate and fairly straight depression trending into a craton commonly from a reentrant adjoining a major basin" (Shatsky, 1964, as quoted in Sengor, 1995, 78).

Rifts that evolve into ocean basins generally are overlain by nascent-ocean and shelf-slope-rise continental margins (Fig. 1.3c–d), whereas fossil rifts adjoining continental margins evolve into reentrants that capture major drainages of continental interiors; major deltas that form at these reentrants commonly construct continental embankments (e.g., Niger Delta) (Dickinson, 1974b; Ingersoll, 1988; Ingersoll and Busby, 1995) (Fig. 1.3f). Upon activation or collision of a continental margin, the rifted-margin sedimentary prisms are intensely deformed, especially at continental promontories (Dewey and Burke, 1974; Graham et al., 1975). As orogeny
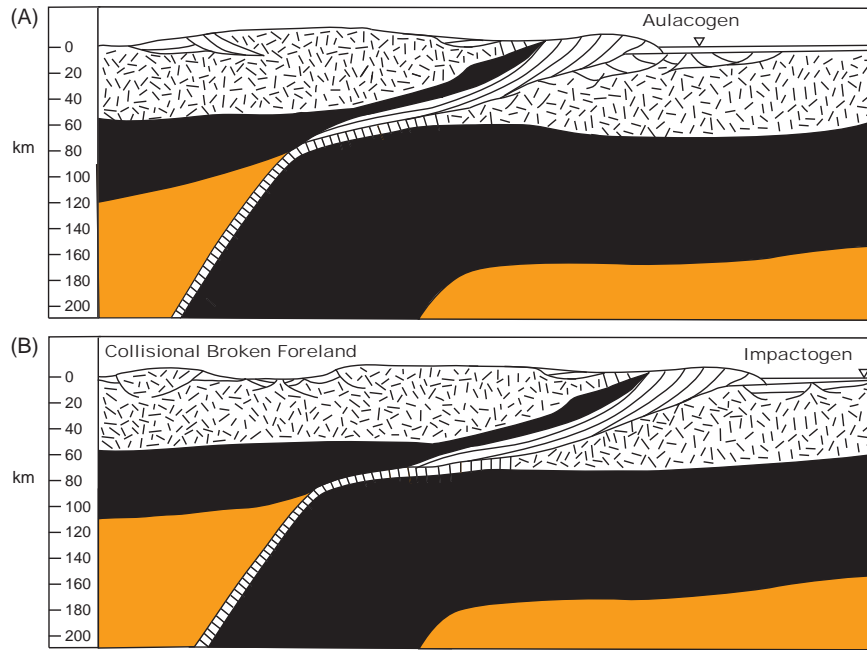
Fig. 1.6. True-scale actualistic analog models for sedimentary basins in continental collisional settings, resulting in hybrid basins. Depiction of aulacogens and impactogens in these cross sections does not show key four-dimensional relations that dictate their histories (see text for discussion). Aulacogens commonly are associated with continental embankments deposited by deltas at the mouths of fossil rifts along intraplate continental margins; therefore, thicker precollisional strata are indicated in (A). Normal faults of the fossil rifts are commonly reactivated during the aulacogen stage (syncollisional). Impactogens do not have precollisional strata or structures to reactivate. The impactogen on the right side of (B) is similar to the proximal proforeland Rhine graben. The collisional broken foreland on the left might include a distal retroforeland impactogen such as the Baikal rift. Symbols same as in Figure 1.3.

proceeds, fossil rifts become aulacogens, which may experience compressional, extensional or translational deformation.

Sengor et al. (1978) and Sengor (1995) developed criteria for distinguishing fossil rifts formed during the opening of nearby oceans that are later closed (aulacogens) from rifts formed due to crustal collision (impactogens). Both types of rift valleys trend at high angles to orogenic belts; however, aulacogens have a rifting history coincident with initiation of a neighboring ocean basin prior to collision, whereas impactogens have no precollisional rift history. Tests for distinguishing them must come from the stratigraphic record because temporal correlation of initial rifting (or lack thereof) is the primary test for their geodynamic origin (Ingersoll and Busby, 1995). Aulacogens tend to form at reentrants along rifted continental margins (Dewey and Burke, 1974), whereas impactogens are more likely to form opposite coastal promontories, where deformation of colliding continents is more intense (Sengor, 1976, 1995). This criterion must be applied cautiously, however, due to the difficulty of definitively reconstructing precollision geometry (e.g., Thomas, 1983, 1985).

Impactogens

Impactogens (Sengor et al., 1978; Sengor, 1995) resemble aulacogens (rifts at high angles to orogenic belts), but without preorogenic stages (Fig. 1.6b). They typically form during attempted subduction of continental crust (during collision, with either another continent or a magmatic arc). Two excellent examples, of contrasting style and tectonic setting, are the middle Cenozoic Rhine graben and the late Cenozoic Baikal rift. The Rhine graben formed as a transtensional impactogen proximal to the Alpine collision orogen (Sengor, 1976). It formed on the subducting plate (Europe), in a proforeland setting. The Baikal rift, which is still active, is also transtensional, but it is distal to the related Himalayan collision (Ingersoll and Busby, 1995). It is part of the collisional broken foreland of central Asia, which is the overriding plate (Fig. 1.6b). Thus, these are end-member types

of impactogens: the Rhine graben formed in a proximal proforeland, whereas the Baikal rift formed in a distal retroforeland. Sengor (1995) discussed these and other examples.

Collisional broken-foreland basins

The collision of continents of varying shapes and sizes usually results in extreme complexity in ancient orogenic belts and related sedimentary basins (e.g., Dewey and Burke, 1974; Graham et al., 1975, 1993; Molnar and Tapponnier, 1975; Sengor, 1976, 1995; Tapponnier et al., 1982). As Tapponnier et al. (1982) demonstrated through the use of plasticine models, the collision of India and Asia has resulted in major intracontinental strike-slip faults, with associated foreland, rift, transtensional, transpressional, and transrotational basins (e.g., Graham et al., 1993; Allen et al., 1999; Yin and Harrison, 2000; Howard et al., 2003). All of these types of basins may form in either proforeland (e.g., Rhine graben) or retroforeland (e.g., Baikal rift) collisional settings.

An excellent ancient example of collisional broken-foreland basins and uplifts is the Pennsylvanian-Permian Ancestral Rocky Mountain (ARM) orogenic belt (e.g., Kluth and Coney, 1981; Kluth, 1986; Dickinson and Lawton, 2003; Blakey, 2008). ARM deformation occurred concurrently with final suturing between Laurasia and Gondwana during the late Carboniferous into the Permian (Graham et al., 1975; Kluth and Coney, 1981; Kluth, 1986; Ingersoll et al., 1995, 2003; Dickinson and Lawton, 2003; Miall, 2008). Foreland and rift basins, commonly with transpressional and transtensional components, respectively, have been documented adjacent to basement-involved uplifts (e.g., Soreghan, 1994; Geslin, 1998; Hoy and Ridgway, 2002; Barbeau, 2003). Reactivation of basement features determined location and character of many ARM uplifts and basins, including reactivation of Proterozoic fossil rifts to form aulacogens (e.g., Sengor, 1995; Marshak et al., 2000; Dickinson and Lawton, 2003; Blakey, 2008; Miall, 2008).

Additional studies of broken forelands include late Proterozoic deformation of North America (e.g., Cannon, 1995), late Paleozoic Appalachian deformation (e.g., McBride and Nelson, 1999; Murphy et al., 1999; Root and Onasch, 1999), late Paleozoic deformation of Europe (e.g., Stollhofen and Stanistreet, 1994; Mattern, 2001; Vanbrabant et al., 2002), and Mesozoic deformation of central Asia (e.g., Sobel, 1999; Vincent and Allen, 1999; Kao et al., 2001; Johnson, 2004; Ritts et al., 2009).

Halokinetic basins

Increased exploration of deep-marine continental margins (especially continental embankments such as the northern Gulf of Mexico) has demonstrated the importance of deformation of salt in producing ponded sedimentary basins (Fig. 1.3f) (Worrall and Snelson, 1989; Winker, 1996; Prather et al., 1998; Badalini et al., 2000; Beaubouef and Friedmann, 2000). Hudec et al. (2009) reviewed subsidence mechanisms for such "minibasins," and suggested that they can be viewed as smaller-scale models of crustal basins. They suggested that subsidence in "minibasins" can be caused by (1) density contrasts (2) diapir shortening (3) extensional diapir fall (4) decay of salt topography (5) sedimentary topographic loading, and (6) subsalt deformation. They also discussed criteria for distinguishing these subsidence mechanisms.

Study of ancient settings, where salt has played important roles in determining kinematic response of weak sediment of contrasting densities to tectonic and gravitational forces (e.g., Giles and Lawton, 2002; Rowan et al., 2003), demonstrates the uniqueness of halokinetic structural development and formation of sedimentary basins. All basins directly related to halokinetic processes (the well-studied "mini-basins" of the Gulf of Mexico, as well as diverse other salt-related basins) are herein termed "halokinetic basins."

Bolide basins

Discovery of an iridium anomaly at the Cretaceous-Paleogene boundary (i.e., Alvarez et al., 1980) raised awareness of the significance of the impact of extraterrestrial objects (bolides) in Earth history. Not only have large impacts resulted in major evolutionary changes (e.g., mass extinctions), but also they have produced widespread sedimentary deposits resulting from tsunamis, landslides, air fall, and other bolide-induced processes (e.g., Bourgeois et al., 1988; Alvarez et al., 1992; Smit et al., 1996; Bralower et al., 1998). The "smoking gun" to explain worldwide bolide-produced sediments at the Cretaceous-Paleogene boundary has been identified as the Chicxulub crater beneath the north coast of Yucatan, Mexico (Hildebrand et al., 1991; Pope et al., 1991). Significant sedimentary basins resulting from filling of impact craters

and related features are herein termed "bolide basins" (Fig. 1.3e).

Bolide basins are now recognized in many localities on Earth: Chicxulub, Chesapeake Bay (Shah et al., 2005; Gohn et al., 2006; Hayden et al., 2008), the North Sea (Stewart and Allen, 2002, 2005), and the Barents Sea (Tsikalas et al., 1998; Dypvik et al., 2004). Detailed stratigraphic and basin analyses have been conducted in some of these basins (e.g., Marin et al., 2001; Parnell et al., 2005; Hayden et al., 2008). Some of these basins are prolific hydrocarbon producers (e.g., Grajalas-Nishimura et al., 2000). An expanding literature concerning bolide basins demonstrates the importance of this type of sedimentary basin (e.g., Glickson and Haines, 2005; Evans et al., 2008). Stewart (2003) discussed criteria for the recognition of bolide basins.

Successor basins

The original definition of successor basins (King, 1966) as "deeply subsiding troughs with limited volcanism associated with rather narrow uplifts, and overlying deformed and intruded eugeosynclines" (Kay, 1951, 107; Eisbacher, 1974) needs modification; "deeply subsiding" and "eugeosynclines" should be replaced by "intermontane" and "terranes," respectively (Ingersoll, 1988; Ingersoll and Busby, 1995). Within the context of plate tectonics, successor basins form primarily in intermontane settings on top of inactive foldthrust belts, suture belts, transform belts, and noncratonal fossil rifts. The presence of successor basins indicates the end of orogenic or taphrogenic activity; therefore, their ages constrain interpretations of timing of suturing, deformation, and rifting (Ingersoll and Busby, 1995). Thus, they have special significance in "terrane analysis"; they represent overlap assemblages which provide minimum ages for terrane accretion (e.g., Howell et al., 1985; Ricketts, 2008).

Little work has been published on actualistic models for such basins; Eisbacher (1974) summarized models based on work on ancient basins in the Canadian Cordillera. This dearth of work may reflect the diversity of successor basins and their tectonic settings. In a sense, all basins are successor basins because they form following some orogenic or taphrogenic event represented in the basement of the basin. In fact, one of Kay's (1951) examples of epieugeosynclines (successor basins) is the post-

Nevadan basin of central California, which is now interpreted as a forearc basin, overprinted in the Cenozoic by transform tectonics (Ingersoll, 1982; Ingersoll and Schweickert, 1986; Dickinson, 1995). Dickinson (1995) discussed examples of "sutural forearc basins," remnants of which are found along suture zones; deposition that occurred following suturing would have been in successor basins (e.g., Ricketts, 2008). Modern use of the term "successor basin" should be restricted to post-orogenic and post-taphrogenic basins that do not fall into any other plate-tectonic framework (Ingersoll and Busby, 1995). For example, most of the southern Basin and Range Province has been tectonically inactive since the Miocene (Wernicke, 1992; Dickinson, 2006). Therefore, modern intermontane basins of this region may be considered successor basins (Ingersoll and Busby, 1995) (Fig. 1.5c).

DISCUSSION

Readers of this review might be overwhelmed by the complexity of tectonic processes controlling the evolution of sedimentary basins, and the resulting complexity of this catalog of basin types. The more we know about these processes and their consequences, the more complex become our models, and the more each basin seems unique (e.g., Dickinson, 1993). This outcome is both exhilarating and frustrating. Exhilaration results from new discoveries of both fact and insight; frustration results from the need to assimilate the overwhelming crush of new information. New models are developed each time insightful simplifications or generalizations are made. Integration of observation, modeling, and experiment is an iterative, self-adjusting process.

The ultimate goal of classifying and reviewing all types of sedimentary basins is the improvement of paleotectonic and paleogeographic reconstructions through the application of actualistic models for basin evolution. Related features, whose recognition aids paleotectonic reconstruction, include suture belts (e.g., Burke et al., 1977), magmatic arcs (e.g., Sengor et al., 1991), foldthrust belts (e.g., McClay, 1992), and metamorphic belts (e.g., Miyashiro, 1973). A skilled basin analyst needs to integrate these topics, as well as geochemistry, geophysics, petrology, paleoecology, and an array of other disciplines. In a complementary manner, workers in these other fields should draw on the insights provided by the sedimentary

record to constrain their paleotectonic reconstructions. I hope that this review encourages this process of interdisciplinary development and testing of models regarding Earth's evolution.

ACKNOWLEDGMENTS

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Part 2

New Techniques and Modeling

Chapter 2

Detrital zircon U-Pb geochronology: current methods and new opportunities

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ABSTRACT

Detrital zircon geochronology is rapidly evolving into a very powerful tool for deter-mining the provenance and maximum depositional age of clastic strata. This rapid evolution is being driven by the increased availability of ion probes and laser ablation inductively coupled plasma (ICP) mass spectrometers, which are able to generate ages efficiently and with sufficient accuracy for most applications. Although large numbers of detrital zircon ages are generated each year, significant uncertainties remain in how data are acquired, which ages are used, how data are plotted, and how age distributions are compared. Improvements in current methods may come from enhanced precision/accuracy of age determinations, better tools for extracting critical information from age spectra, abilities to determine other types of information (e.g., Hf, O, Li isotope signatures, rare earth element (REE) patterns, cooling ages, structural information) from the dated grains, and availability of a database that provides access to detrital zircon age determinations from sedimentary sequences around the world.

Keywords: zircon; geochronology; detrital; provenance; analytical methods

INTRODUCTION

Detrital zircon geochronology has evolved rapidly during the past ∼20 years, from a technique with apparently limited application to a nearly indis-pensible method of investigating sedimentary units and their source regions. This evolution, referred to informally as the "DZ revolution," is largely a result of technical developments that have allowed researchers to efficiently determine U-Pb ages on individual zircon crystals (see Davis et al. [2003] and Kosler and Sylvester [2003] for excellent reviews of this history). Prior to these developments, most detrital zircon analyses were conducted on multigrain fractions. The prevailing strategy was to analyze groups of grains with similar characteristics (e.g., color, shape, round-ing, etc.) in hopes that the grains in each fraction were of the same age (e.g., LeDent et al., 1964; Hart and Davis, 1969; Girty and Wardlaw; 1984; Erdmer and Baadsgaard, 1987; Gehrels et al., 1990; Ross and Bowring, 1990). While sufficient

to determine general age distributions, this pro-cedure was generally inadequate to identify spe-cific source ages.

Analysis of individual zircon crystals became feasible during the 1980s and early 1990s utilizing both ion probes (e.g., Froude et al., 1983; Dodson et al., 1988; Ireland, 1992) and isotope dilution-thermal ionization mass spectrometers (ID-TIMS) (e.g., Davis et al., 1989). Technical developments that allowed for ID-TIMS analyses of individual crystals included preparation of a ^{205}Pb spike (Krogh and Davis, 1975; Parrish and Krogh, 1987) and design of Teflon microcapsules for low-blank grain dissolution (Parrish, 1987). During the late 1980s and early 1990s, ion probes and ID-TIMS techniques became increasingly used for detrital zircon geochronologic studies. A dramatic increase in detrital zircon geochronology occurred in the late 1990s through early 2000s, when laser-ablation ICP mass spectrometers (LA-ICPMS) were devel-oped to the point that U-Pb analyses could be conducted in a robust fashion (Fryer et al., 1993;

Machado and Gauthier, 1996; Horn et al., 2000; Kosler et al., 2001; Li et al., 2001; Machado and Simonetti, 2001; Horstwood et al., 2003).

At present, detrital zircon geochronology is used primarily for four principle applications (Fedo et al., 2003; Anderson, 2005):

- Provenance studies, where ages of detrital minerals are compared with ages of potential source terranes to determine ultimate sources of sediment
- Where provenance is known, determination of ages and characteristics of rocks in source terranes
- Correlation of sedimentary units, where ages of detrital minerals are compared in an effort to evaluate possible linkages between different sedimentary units
- Maximum depositional age, where the youngest age component in a clastic unit provides the earliest possible age of deposition

Although our research community is currently experiencing an explosion in the number of detrital zircon analyses that are conducted and published, it is disconcerting that we are not yet able to answer some basic questions about how to collect, display, and interpret U-Pb geochronologic data applied to detrital minerals. Some of these issues have been highlighted recently by Horstwood et al. (2009). Examples include the following:

- What is the optimal instrumentation used for a detrital zircon study?
- Which ages should be used, and how should these ages be evaluated and filtered?
- How many analyses should be conducted from each sample, and how should grains be selected for analysis?
- What is the most effective way to display detrital zircon data – Pb/U Concordia diagrams, Terra-Wasserburg diagrams, histograms, age-distribution diagrams, or cumulative probability plots?
- What is the best method of describing a set of detrital zircon ages (e.g., the youngest age component)?
- What is the best method for comparing age distributions of several samples?

This chapter is a progress report on attempts to address these questions, with an emphasis on current methods for gathering and interpreting U-Pb information and a brief look forward to future opportunities of gathering critical information from detrital minerals. Readers are referred to Fedo et al. (2003) for an excellent summary of the history and applications of detrital zircon geochronology, and Gehrels (2000), Anderson (2005), and Nemchin and Cawood (2005) for a discussion of statistical analysis of detrital zircon data.

WHAT IS THE OPTIMAL INSTRUMENTATION USED FOR A DETRITAL ZIRCON STUDY?

Three types of instruments are routinely used for determining U-Pb ages of detrital zircons. Following is a brief outline of the analytical methods used with each type of instrument, together with an evaluation of strengths and weaknesses for conducting U-Pb analyses of detrital zircons.

Isotope dilution-thermal ionization mass spectrometry (ID-TIMS)

ID-TIMS analyses require dissolution of complete crystals or portions of crystals, addition of an isotopic tracer (commonly ^{205}Pb and ^{233}U), chemical separation of U and Pb, and isotopic analysis by TIMS (Bowring and Schmitz, 2003; Parrish and Noble, 2003; Mattinson, 2005). The chemical dissolution and separation portion of the analysis is quite time consuming and needs to be conducted in an ultra-clean environment to reduce contaminant Pb and U, but results in a very pure analyte that yields high-precision ($\sim$0.1% at 2-sigma) isotope ratios and Pb/U concentrations. U-Pb ages determined by ID-TIMS are accordingly of the best possible precision and accuracy, and are essential for applications that require high temporal resolution. In most cases, however, this high precision is not necessary for a detrital zircon study.

Secondary ion mass spectrometry (SIMS or ion probe)

SIMS analyses are generally conducted on the polished surface of a crystal that has been mounted in epoxy along with "standards," which are crystals of the same mineral that are of known age and isotopic composition (Ireland and Williams, 2003). Because sputtering takes place at low temperature and high vacuum, backgrounds of Pb and U are

low, and analyses can be conducted on very small volumes ($\sim 1\,ng$) of material. This provides opportunities to determine U-Pb ages on small portions of crystals, with typical pits that are 10–30 microns diameter and ~ 1 micron depth. Because it is not possible to add an isotopic tracer, ages are corrected for instrumental fractionation by standard-sample bracketing (alternating between standards and unknowns, correcting the standards to the known age, and applying the same correction factor to the unknowns). This method yields ages with a precision and accuracy of 1–2% (2-sigma). Isotopic measurements are conducted by sequential analysis of U-Th-Pb (and other) isotopes, which requires a typical analysis time of ~ 15 minutes. This technique is ideal for studies that require high spatial resolution (especially in the depth dimension), such as analysis of complex zircon crystals. Ion probes are also able to analyze other elements at the same time that U-Pb ages are determined, so it is now possible to also characterize detrital minerals for Ti and Zr concentrations, REE concentrations, and oxygen isotopes for constraints on petrogenesis (e.g., Mojzsis et al., 2001; Valley, 2003; Wooden et al., 2007).

Laser-ablation inductively coupled plasma mass spectrometry (LA-ICPMS)

LA-ICPMS methodology is similar to SIMS in that fractionation is determined by standard-sample bracketing, and that analyses are conducted on a polished crystal surface (e.g., Kosler and Sylvester, 2003). This method also yields ages with a precision and accuracy of 1–2% (2-sigma; Machado and Simonetti, 2001; Horstwood et al., 2003; Kosler and Sylvester, 2003; Chang et al., 2006; Gehrels et al., 2008; Horstwood, 2008). An advantage of LA-ICPMS is their much faster analysis time, facilitated with some instruments by having sufficient dispersion and enough collectors to be able to measure U and Pb simultaneously. A disadvantage, however, is that plasma ionization involves high flow rates of Ar gas, at atmospheric pressure, and at high temperature, all of which result in high background counts of Pb and Hg (which interferes with ^{204}Pb). Achieving high signal:background requires a fast rate of ablation and in most cases a larger volume of analyzed material (pits are typically 30 micron diameter by 10–20 micron depth). This faster rate of ablation makes analysis by LA-ICPMS very efficient, with typical analysis times of several minutes.

LA-ICPMS instruments are also ideally suited to analysis of other elements such as Hf-Lu-Yb (for Hf isotope determinations) and trace/rare earth elements (e.g., Machado and Simonetti, 2001; Woodhead et al., 2004; Gerdes and Zeh, 2006; Flowerdew et al., 2007; Mueller et al., 2007; Yuan et al., 2008, Kemp et al., 2009).

WHICH AGES SHOULD BE USED, AND HOW SHOULD AGES BE EVALUATED AND FILTERED?

The U-Pb system is particularly powerful for geochronology because (1) there are two decay systems ($^{238}U \rightarrow {}^{206}Pb$ and $^{235}U \rightarrow {}^{207}Pb$), (2) half lives for the two systems are appropriate for use through all but the most recent portion of Earth time, (3) the two decay systems are linked because $^{238}U/^{235}U$ is constant (137.88; Steiger and Jäger, 1977) in nearly all crustal rocks, and (4) there is a non-radiogenic isotope of Pb (^{204}Pb) that can be used to account for Pb present in the crystal at the time of formation. These aspects allow graphical representation of U-Pb ages on a Pb*/U Concordia diagram (Wetherill, 1956; Fig. 2.1), which plots $^{206}Pb^*/^{238}U$ versus $^{207}Pb^*/^{235}U$ as a function of age (* indicates that initial Pb has been subtracted). The Concordia diagram also shows $^{206}Pb^*/^{207}Pb^*$, which is the slope of a line from the origin through the analysis ($^{206}Pb^*/^{207}Pb^* = {}^{206}Pb^*/^{238}U/[^{207}Pb^*/^{235}U \times 137.88]$) (Fig. 2.1).

Operationally, $^{206}Pb^*/^{238}U$ is determined by comparison with a tracer solution containing known amounts of Pb and U (e.g., ^{205}Pb and ^{233}U) for ID-TIMS, or by comparison with standards for SIMS and LA-ICPMS. $^{206}Pb^*/^{207}Pb^*$ generally requires only a minor correction because there is little instrumental fractionation of ^{206}Pb relative to ^{207}Pb. $^{207}Pb^*/^{235}U$ is usually not measured directly – instead this ratio is calculated from measured $^{206}Pb^*/^{238}U$, measured $^{206}Pb^*/^{207}Pb^*$, and known $^{238}U/^{235}U$. Because ^{235}U is much (137.88 times) smaller than ^{238}U, measurement of ^{235}U would add significant uncertainty to the $^{207}Pb^*/^{235}U$ age. An analysis is plotted on a Concordia diagram based on $^{206}Pb^*/^{207}Pb^*$ (expressed as $^{207}Pb^*/^{206}Pb^*$ in many labs) and $^{206}Pb^*/^{238}U$. The three available ages are then shown as the intersection of $^{206}Pb^*/^{238}U$, $^{207}Pb^*/^{235}U$, and $^{206}Pb^*/^{207}Pb^*$ lines with Concordia (Fig. 2.1).

Uncertainties for $^{206}Pb^*/^{238}U$, $^{207}Pb^*/^{235}U$, and $^{206}Pb^*/^{207}Pb^*$ generally form a zircon-shaped(!)
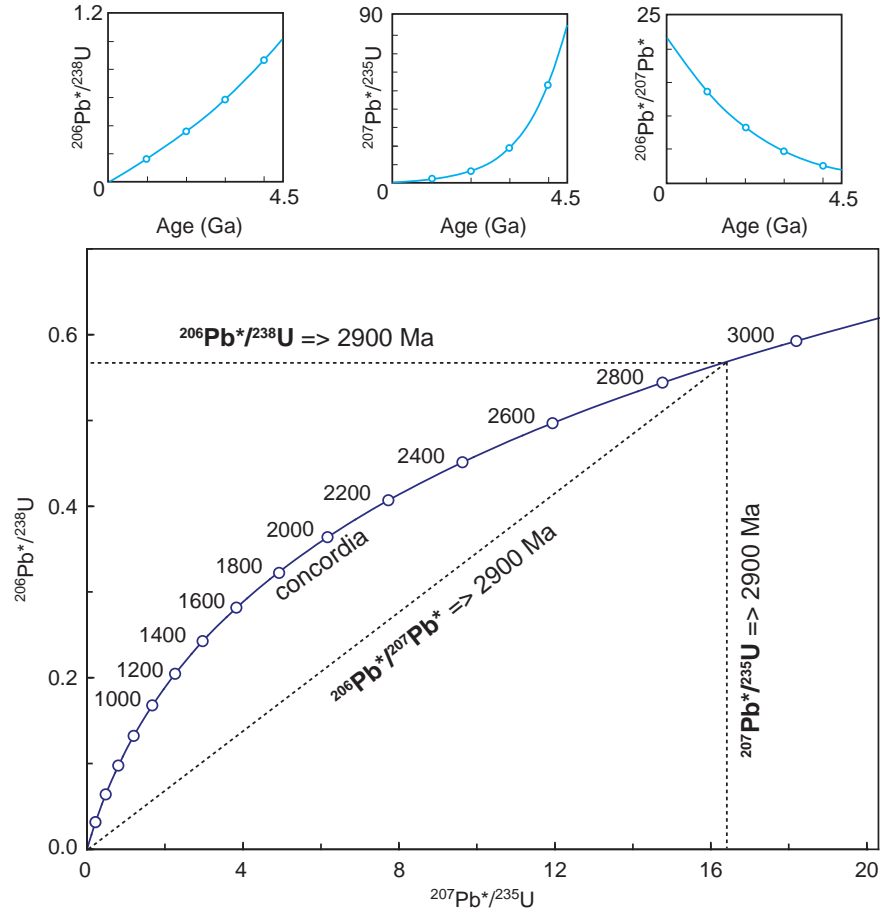
Fig. 2.1. Pb*/U Concordia diagram, showing the three chronometers that are commonly used. The Concordia line shows coincidence of the three chronometers as a function of age (Wetherill, 1956). Note that ^{206}Pb*/^{238}U and ^{206}Pb*/^{207}Pb* are the two independent measurements; ^{207}Pb*/^{235}U is calculated from these ratios and the modern ^{238}U/^{235}U (137.88).

polygon, which is commonly expressed as a continuous probability density function (e.g., Ludwig, 2008). The relative values of the three uncertainties vary as a function of age: for "young" samples, ^{206}Pb*/^{238}U age is the most precise and ^{206}Pb*/^{207}Pb* age is the least precise, whereas for "old" samples the values are reversed, with ^{206}Pb*/^{207}Pb* age more precise than ^{206}Pb*/^{238}U age (Fig. 2.2). The uncertainty of ^{207}Pb*/^{235}U (in age) is always intermediate in value. In theory and in practice, all three uncertainties have the same value at about 1.4 Ga (Gehrels, 2000; Nemchin and Cawood, 2005; Gehrels et al., 2008). Because of these variations, the age resolution of the U-Pb system is poorest at ~1.4 Ga, with improving precision of ^{206}Pb*/^{238}U age for younger ages and improving precision of ^{206}Pb*/^{207}Pb* age for older ages.

If all three ages (^{206}Pb*/^{238}U, ^{207}Pb*/^{235}U, and ^{206}Pb*/^{207}Pb*) are similar within error, an analysis plots on the Concordia line and is described as being "concordant" (Wetherill, 1956). Unfortunately, it is common for an analysis to lie below Concordia, in which case it is referred to as being "discordant." This results in ages that increase from ^{206}Pb*/^{238}U to ^{207}Pb*/^{235}U to ^{206}Pb*/^{207}Pb*. Discordance most commonly occurs due to loss of Pb during younger thermal/hydrothermal activity (in which case the analysis moves down the ^{206}Pb*/^{207}Pb* line that existed at the time of disturbance), or due to inheritance of older material (in which the analysis moves up along a mixing line). "Reverse discordant" analyses, which plot above Concordia, are rare, and most likely result from inaccurate measurement of ^{206}Pb*/^{238}U. The degree of discordance is best expressed as a percentage of the ^{206}Pb*/^{238}U age divided by the ^{206}Pb*/^{207}Pb* age, with a perfectly concordant analysis having 0% discordance (discordance $= 100 - 100 \times [^{206}$Pb*/^{238}U age/^{206}Pb*/^{207}Pb* age]. Conversely, this ratio can be expressed
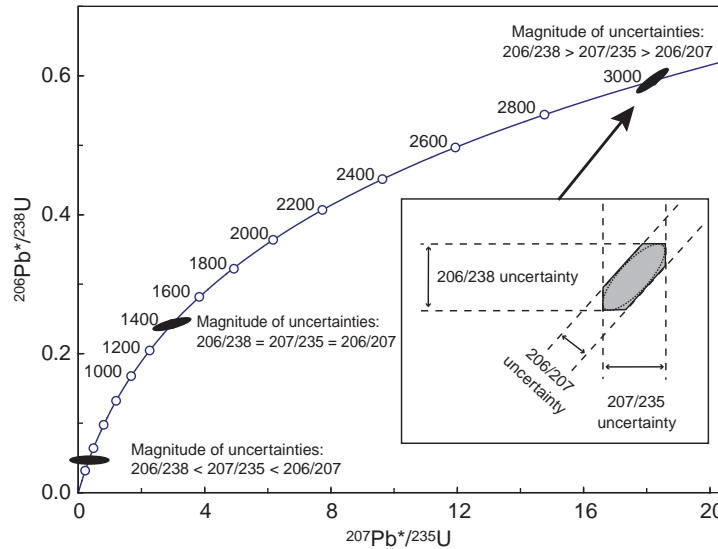
Fig. 2.2. Example showing the orientation and magnitude of uncertainty ellipses for Phanerozoic, Proterozoic, and Archean ages. The relative magnitudes (in millions of years) of the three primary uncertainties are also indicated. Inset on right is a graphical depiction of the three primary uncertainties.

as degree of concordance, with a perfectly concordant analysis equal to 100% (concordance = 100× [^{206}Pb*/^{238}U age/^{206}Pb*/^{207}Pb* age.

Because zircons are more susceptible to Pb loss with increasing age (due to increasing lattice damage during radioactive decay), it is common for Precambrian zircons to be discordant by a few percent to a few tens of percent. For example, Figure 2.3 is an example of a zircon that crystallized at ~1500 Ma and experienced Pb loss at ~500 Ma. The ^{206}Pb*/^{238}U, ^{207}Pb*/^{235}U, and ^{206}Pb*/^{207}Pb* ages are all younger than the crystallization age, with the ^{206}Pb*/^{207}Pb* age the closest to the true age. Because of this, for reasonable (~10–30%) degrees of discordance it is generally more accurate to use ^{206}Pb*/^{207}Pb* ages for zircons that are younger than 1.4 Ga, even though the ^{206}Pb*/^{238}U age is more precise. A common cutoff is 0.8–1.0 Ga (Gehrels, 2000; Gehrels et al., 2008), which is a balance between the more accurate ^{206}Pb*/^{207}Pb* age (assuming that discordance is due to Pb loss) and the more precise ^{206}Pb*/^{238}U age.

For analyses that are discordant due to Pb loss, it is important to realize that even the ^{206}Pb*/^{207}Pb* age is younger than the true crystallization age (Fig. 2.3). The inaccuracy of the ^{206}Pb*/^{207}Pb* age becomes worse as discordance increases, and as the age of the Pb-loss event increases, even though precision of the analysis does not vary. The ^{206}Pb*/^{238}U, ^{207}Pb*/^{235}U, and ^{206}Pb*/^{207}Pb* ages are also inaccurate if discordance results from inheritance, with potential inaccuracy increasing with increasing discordance. It is accordingly

common in a detrital zircon study to filter data based on degree of discordance. But what is the appropriate cutoff to use for discordance?

Unfortunately, the appropriate level of discordance filter needs to be determined for each data set in light of the goals of the study and the complexities encountered. For example, if a study yields a mix of Phanerozoic and Archean ages, and the relative proportions of these ages are important, a generous (e.g., 30%) discordance cutoff might be appropriate so that most Precambrian ages are retained. If instead the main objective of a study is to test for a specific Late Archean age, it would be appropriate to use a tight discordance filter (e.g., 10%), or even use only concordant data, to achieve the best possible age resolution. And last, if a study yielded only young ages (e.g., <100 Ma), a discordance filter could not even be applied reliably because of the difficulty of determining reliable ^{206}Pb*/^{207}Pb* ages, and hence degree of discordance, for young ages.

A discordance filter is also not assured of yielding robust ages because Pb loss soon after crystallization and/or inheritance of slightly older components will yield analyses that are analytically concordant but of inaccurate age. An alternative method is to place significance primarily on analyses that belong to a cluster, given that Pb loss and inheritance always scatter analyses away from their true age, A reasonable methodology is to filter for clustering, with a cluster defined, for example, as three or more analyses that overlap at 2-sigma uncertainty.