A U–Pb and $^{40}Ar/^{39}Ar$ time scale for the Jurassic

J. Pálfy, P.L. Smith, and J.K. Mortensen

Abstract: Published time scales provide discrepant age estimates for Jurassic stage boundaries and carry large uncertainties. The U–Pb or $^{40}Ar/^{39}Ar$ dating of volcaniclastic rocks with precisely known stratigraphic age is the preferred method to improve the calibration. A radiometric age database consisting of fifty U–Pb and $^{40}Ar/^{39}Ar$ ages was compiled to construct a revised Jurassic time scale. Accepted ages have a precision of ±5 Ma (2σ) or better and are confined to no more than two adjacent stages. The majority of these calibration points result from integrated bio- and geochronologic dating in the western North American Cordillera and have not been previously used in time scales. Direct dates are available only for the Triassic–Jurassic boundary and the initial boundary of the Crassicosta chron and the Callovian stage. The chronogram method was used to estimate all Early and early Middle Jurassic zone boundaries (attempted here for the first time), late Middle Jurassic substage boundaries, and Late Jurassic stage boundaries. Significant improvement is achieved for the Pliensbachian and Toarcian, where six consecutive zone boundaries are determined. The derived zonal durations are disparate, varying between 0.4 and 1.6 Ma. The latest Jurassic isotopic database remains too sparse, therefore chronogram estimates are improved using interpolation based on magnetochronology. The initial boundaries of Jurassic stages are proposed as follows: Berriasian (Jurassic–Cretaceous): $141.8^{+2.5}_{-1.8}$ Ma; Tithonian: $150.5^{+3.4}_{-2.8}$ Ma; Kimmeridgian: $154.7^{+3.8}_{-3.3}$ Ma; Oxfordian: $156.5^{+3.1}_{-5.1}$ Ma; Callovian: $160.4^{+1.1}_{-0.5}$ Ma; Bathonian: $166.0^{+3.8}_{-5.6}$ Ma; Bajocian: $174.0^{+1.2}_{-7.9}$ Ma; Aalenian: $178.0^{+1.0}_{-1.5}$ Ma; Toarcian: $183.6^{+1.7}_{-1.1}$ Ma; Pliensbachian: $191.5^{+1.9}_{-4.7}$ Ma; Sinemurian: $196.5^{+1.7}_{-5.7}$ Ma; Hettangian (Triassic–Jurassic): 199.6 ± 0.4 Ma.

Résumé : Les échelles de temps publiées donnent des estimations d'âges divergentes pour les limites des étages du Jurassique; elles comportent aussi de grandes incertitudes. Les datations U–Pb ou $^{40}Ar/^{39}Ar$ de roches volcanoclastiques dont l'âge stratigraphique est connu avec précision est la méthode de choix pour améliorer la calibration. Une base de données d'âges radiométriques contenant cinquante âges U–Pb et $^{40}Ar/^{39}Ar$ a été compilée afin de construire une échelle de temps révisée pour le Jurassique. Les âges acceptés ont une précision de ± 5Ma (2σ) ou mieux et ils portent au plus sur deux étages subséquents. La plupart de ces points de calibration proviennent d'une intégration de datations bio- et géochronologiques de l'ouest de la Cordillère nord-américaine et ils n'ont pas servi antérieurement pour des échelles de temps. Des dates directes ne sont disponibles que pour la limite Trias-Jurassique et la limite initiale du chron Crassicosta et de l'étage Callovien. La méthode du chronogramme a été utilisée pour estimer toutes les limites des zones du Jurassique précoce et moyen précoce (il s'agit ici d'un premier essai), pour les limites des sous-étages du Jurassique moyen tardif et les limites des étages du Jurassique tardif. Le Pliensbachien et le Toarcien ont connu une amélioration importante; les limites de six zones consécutives y ont été déterminées. Les durées des zones qu'on en a tirées sont disparates et varient entre 0,4 et 1,6 Ma. La base de données isotopiques pour le Jurassique tardif demeure trop clairsemée; les estimations tirées de chronogrammes sont donc meilleures en utilisant une interpolation basée sur la magnétochronologie.On propose les limites initiales suivantes pour les étages du Jurassique : Berriasien (Jurassique–Crétacé) : $141{,}8^{+2{,}5}_{-1{,}8}$ Ma; Tithonien : $150{,}5^{+3{,}4}_{-2{,}8}$ Ma; Kimmeridgien : $154{,}7^{+3{,}8}_{-3{,}3}$ Ma; Oxfordien : $156{,}5^{+3{,}1}_{-5{,}1}$ Ma; Callovien : $160{,}4^{+1{,}1}_{-0{,}5}$; Bathonien : $166{,}0^{+3{,}8}_{-5{,}6}$ Ma; Bajocien : $174{,}0^{+1{,}2}_{-7{,}9}$ Ma; $178{,}0^{+1{,}0}_{-1{,}5}$ Ma; Toarcien : $183{,}6^{+1{,}7}_{-1{,}1}$ Ma; Pliensbachien : $191{,}5^{+1{,}9}_{-4{,}7}$ Ma; Sinémurien : $196{,}5^{+1{,}7}_{-5{,}7}$ Ma; Hettangien (Trias-Jurassique) :199,6 ± 0,4 Ma.

[Traduit par la Rédaction]

Introduction

Geochronologic scales or time scales express the estimated numerical ages of chronostratigraphic units in millions of years. Jurassic chronostratigraphic units are defined on ammonite biochronology based on the well established zonal standard from northwest Europe. In contrast to the high resolution of biochronology, the Jurassic time scale remains less well calibrated than most other periods. Conflicting and imprecise estimates of the different Jurassic time scales result from the scarcity of biochronologically well constrained isotopic ages and the preponderance of low-temperature K–Ar and Rb–Sr dates of poor accuracy and precision (Pálfy 1995). In the North American Cordillera, systematic effort was made to generate new calibration points by integrating ammonite biochronology of marine sediments and U–Pb zircon dating of interbedded volcanic or volcaniclastic rocks. The result is 18 new calibration points (Pálfy et al. 1997, 1999, 2000*a*, 2000*b*). Additional radiometric ages, also useful for time-scale calibration, have been reported in other recent studies. The revised Jurassic time scale presented here differs from the previous ones be-

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cause it (1) is more selective in database compilation by employing high precision U–Pb and $^{40}Ar/^{39}Ar$ ages only; (2) uses zonal level biochronology and attempts to estimate chron boundary ages; (3) rejects scaling based on the assumption of equal duration of biochronologic units and minimizes the use of interpolation.

The proposed time scale is compared with previous ones, of which the most frequently cited scales are abbreviated as follows: NDS, Numerical dating in stratigraphy (Odin 1982); DNAG, Decade of North American Geology (Kent and Gradstein 1985; Palmer 1983); EXX88, EXXON (Haq et al. 1988); GTS89, Geologic time scale 1989 (Harland et al. 1990), OD94, Odin (Odin 1994); and MTS94, Mesozoic time scale (Gradstein et al. 1994, 1995).

Methods

The revised Jurassic time scale is the result of combining methods that were successfully employed in previous work with new approaches made possible by recent advances in geochronometry and biochronology. Typically, previous time scale calibrations have utilized three different aproaches (Odin 1994): (1) manual construction that considers each relevant isotopic date individually and weighs them subjectively to arrive at best boundary estimates (as in OD94); (2) statistically oriented methods that treat each accepted date equally and derive the boundary estimates mathematically (as in GTS89 and MTS94); and (3) reliance on a select set of a few dates judged most reliable and determination of the intervening boundaries by interpolation, assuming either an equal duration of biochronologic units or a constant spreading rate deduced from oceanic magnetic anomalies (e.g., EXX88). Some combination of the above is more common in recent works (GTS89, MTS94). Manual construction lacks rigorous methodology and reproducibility but offers flexibility. This method would suffice if each boundary had available isotopic age constraints, a situation not yet attainable for the Jurassic. Statistical methods offer the advantage of handling large numbers of dates efficiently and providing reproducible results. Two methods are tested and available: the chronogram method (Harland et al. 1990) and the maximum likelihood method (Agterberg 1988). They are similar in assuming a random distribution of isotopic ages and their results converge for densely sampled intervals (Agterberg 1988). We prefer to use the semi-rigorous chronogram method that can accommodate, after a slight modification, isotopic ages with asymmetric error that are common among the U–Pb ages.

In previous time scales, interpolation arose from the sparseness of available isotopic ages. The combination of magnetochronology and biochronology provides a powerful tool that allows the interpolation of boundaries, if some magnetochrons are directly dated isotopically and a constant spreading rate is assumed during the formation of oceanic magnetic anomalies (Hailwood 1989). This assumption appears to be tenable for the Late Jurassic (Channell et al. 1995). As no oceanic crust older than Callovian is preserved, the method is not applicable for the Early and Middle Jurassic. In that interval, most scales use some form of biochronologically based interpolation, assuming equal duration of chrons or subchrons (Gradstein et al. 1994; Harland et al. 1990). This method is inadequate based on three independent lines of evidence that suggest widely disparate durations for Mesozoic ammonite zones: (1) direct high-resolution isotopic dating of Triassic (Mundil et al. 1996) and Cretaceous (Obradovich 1993) ammonite zones; (2) Milankovitch cyclostratigraphy of Jurassic (Smith 1990) and Cretaceous (Gale 1995) ammonite zones; and (3) the width of oceanic magnetic anomalies calibrated to Late Jurassic ammonite zones (Ogg and Gutowski 1996; Ogg et al. 1991).

The need for interpolation is eliminated if reliable isotopic ages are available from each zone or stage. The most promising new development in time-scale studies is the use of high-precision U–Pb and $^{40}Ar/^{39}Ar$ dating on volcanic flows and volcaniclastic layers from biochronologically well dated sections (e.g., Mundil et al. 1996; Obradovich 1993). The present study uses this approach in a systematic effort to generate critical U–Pb ages for the Jurassic (Pálfy et al. 1997, 1999, 2000*a*, 2000*b* in press). For concordant analyses, the preferred interpreted ages are based on the calculated $^{206}Pb/^{238}U$ ages and their errors (Ludwig 1998). Less precise and reliable is the use of $^{207}Pb/^{206}Pb$ ages or concordia intercept ages, employed for discordant data. In the lack of duplicate concordant analyses, the interpreted age often has asymmetric errors, where the lower limit is derived from the error bound of the $^{206}Pb/^{238}U$ age of the most concordant fraction and the upper limit is taken from the error bound of the mean $^{207}Pb/^{206}Pb$ age. Such conservative error assignment is warranted for analyses of multigrain fractions, which are inherently vulnerable to averaging mild Pb loss and subtle inheritance.

A conservative, all-inclusive database compilation is advocated in GTS89 "… that does not exclude any generally accepted data [and therefore] introduces considerable stability into the time scale" (Harland et al. 1990). We note that GTS89 and MTS94 include K–Ar ages that were produced as early as 1959, whereas there is an emerging consensus among geochronologists that U–Pb and $^{40}Ar/^{39}Ar$ systems are the most reliable and precise geochronometers currently available. We chose to emphasize accuracy over stability, therefore we give priority to U–Pb and $^{40}Ar/^{39}Ar$ ages and omit ages derived from materials with low closure temperature. Such an approach was already pursued for the Cretaceous (Kowallis et al. 1995; Obradovich 1993).

A valid comparison of U–Pb and $^{40}Ar/^{39}Ar$ ages requires the use of external errors that take into account the decay constant uncertainties (Renne et al. 1998a). One of the weaknesses of the earlier time scales was the lack of consideration of this source of systematic error, which is not routinely published. In addition to the analytical error, the decay constant uncertainty of $^{206}Pb/^{238}U$ ages is ±0.2 Ma and of $^{207}Pb/^{206}Pb$ ages is ±4.0 Ma for samples of Jurassic age. External errors of $^{40}Ar/^{39}Ar$ ages arise from interlaboratory calibration of standards and the decay constant uncertainties, which significantly exceed those of the uranium (Renne et al. 1998b).

Chronostratigraphic framework

The Jurassic chronostratigraphy is traditionally based on the sequence of northwest European ammonite faunas. A well established, hierarchic scheme of stages, zones, subzones and, for many intervals, horizons has been developed, although among the Jurassic stage boundaries, only the base of the Bajocian has been formally defined by means of Global Boundary Stratotype Section and Point (Pavia and Enay 1997). Nevertheless, considerable consensus exists regarding the placement of most boundaries.

This study relies heavily on North American data, mainly because the European successions generally lack isotopically datable horizons. It is, therefore, practical to use a regional ammonite biochronologic scheme, which is readily applicable to constrain the majority of the isotopic dates and correlatable with the northwest European standard. Such North American ammonite biochronologic standards have recently been developed for the Pliensbachian (Smith et al. 1988), Toarcian (Jakobs et al. 1994), Aalenian (Poulton and Tipper 1991), and Bajocian (Hall and Westermann 1980; Hillebrandt et al. 1992) stages. Well documented local zonations also exist for the Hettangian (Tipper and Guex 1994) and Sinemurian (Pálfy et al. 1994); because their non-standard units (i.e., informally defined assemblages) have proved to have widespread applicability, herein we treat them equivalent to zones (chrons).

Correlation to the northwest European standard zonation was carefully considered by the respective authors of North American stage zonations and is compiled here in Fig. 1. A case study of the chronostratigraphic error introduced by interregional correlation demonstrates it to be no more than one subchron (Pálfy et al. 1997).

Endemism of ammonite faunas increases from the Late Bajocian onward. The Bathonian through Oxfordian faunal succession is increasingly well understood (Callomon 1984; Poulton et al. 1994), even though no formal regional zonation has been proposed. Precise correlation, however, is hampered by significant differences between North American and European faunas, therefore, only substage-level subdivision is applied here. The Kimmeridgian and Tithonian are not subdivided here due to the general scarcity of ammonites in North America.

The isotopic age database

Our database of critical isotopic ages is derived from three sources: (1) U–Pb ages from the North American Cordillera (either produced as part of this project or obtained by other workers and the biochronologic constraints critically reviewed and (or) revised by us); (2) U–Pb or $^{40}Ar/^{39}Ar$ ages culled from the databases of previous time scales (mainly from GTS89 and MTS94); and (3) recently reported U–Pb or $^{40}Ar/^{39}Ar$ ages from outside the Cordillera that have not been used in time scale calibration before. Only ages with adequately documented analytical data are included (i.e., dates appearing only in abstracts are not included), with the exception of some recently obtained Cordilleran ages that were made available by our colleagues through personal communications and are currently being prepared for publication.

The isotopic ages were screened for accuracy, precision, and quality of chronostratigraphic constraints. Accuracy is adequate if reproducibility is demonstrated and (or) the error assignment is conservative. A few unresolved dates that ex-

Fig. 1. Early and Middle Jurassic ammonite biochronological units of North America and their correlation with the northwest European standard. Compiled from published North American regional zonations (Hall and Westermann 1980; Hillebrandt et al. 1992; Jakobs et al. 1994; Pálfy et al. 1994; Poulton and Tipper 1991; Smith et al. 1988; Tipper and Guex 1994).

STAGE		NORTH AMERICAN AMMONITE CHRONS	NW EUROPEAN AMMONITE CHRONS
BAJOCIAN	L	Epizigzagiceras	Parkinsoni
			Garantiana
		Rotundum	Subfurcatum
	E	Oblatum	Humphresianum
		Kirschneri	
			Sauzei
		Crassicostatus	
		Widebayense	Laeviuscula
			Discites
AALENIAN		Howelli	Concavum
			Murchisonae
		Scissum	
		Westermanni	Opalinum
TOARCIAN	L	Yakounensis	Levesquei
		Hillebrandti	Thouarsense
	M	Crassicosta	Variabilis
		Planulata	Bifrons
	E	Kanense	Falciferum
			Tenuicostatum
PLIENSBACHIAN	L	Carlottense	Spinatum
		Kunae	Margaritatus
			Davoei
	E	Freboldi	
		Whiteavesi	Ibex
		Imlayi	Jamesoni
		Tetraspidoceras	
SINEMURIAN	L		Raricostatum
		Plesechioceras? harbledownense	Oxynotum
		Asteroceras varians	Obtusum
	E	Arnioceras arnouldi	Turneri
			Semicostatum
		Coroniceras	Bucklandi
HETTANGIAN		Canadensis	
		Pseudaetomoceras	Angulata
		Franziceras	Liasicus
		Euphyllites	Planorbis
		Psiloceras	

Table 1. Listing of selected critical isotopic ages.

Item	Short name	Age (Ma)	Error: Error Int. +2σ	Error: Error Int. –2σ	Error: Ext. +2σ	Error: Ext. –2σ	Base[a]	Top[a]	Method	Reference
1	Guichon Creek batholith	210	3	3	3	3	20	30	U–Pb[e]	Mortimer et al. 1990
2	Griffith Creek sill	205.8	1.5	3.1	4.3	3.1	20	30	U–Pb[h]	Thorkelson et al. 1995
3	Griffith Creek flow	205.8	0.9	0.9	4.1	4.1	20	30	U–Pb[g]	Thorkelson et al. 1995
4	Red Mtn.: Biotite porphyry	201.8	0.5	0.5	0.5	0.5	22	32	U–Pb[e]	Greig et al. 1995
5	Kunga Island	199.6	0.3	0.3	0.4	0.4	32	32	U–Pb[d]	Pálfy et al. 2000*a*
6	North Mtn. basalt	201.7	1.4	1.1	1.4	1.1	30	40	U–Pb[e]	Hodych and Dunning 1992
7	Gettysburg sill	201.3	1.0	1.0	1.0	1.0	30	40	U–Pb[e]	Dunning and Hodych 1990
8	Palisades sill	200.9	1.0	1.0	1.0	1.0	30	40	U–Pb[e]	Dunning and Hodych 1990
9	Puale Bay 1	200.4	2.7	2.8	5.0	2.8	42	42	U–Pb[h]	Pálfy et al. 1999
10	Puale Bay 2	197.8	1.0	1	1.0	1.0	42	43	U–Pb[e]	Pálfy et al. 1999
11	Puale Bay 3	197.8	1.2	0.4	1.2	0.4	42	43	U–Pb[e]	Pálfy et al. 1999
12	Cambria Icefield	199	2	2	2	2	40	40	U–Pb[e]	Greig and Gehrels 1995
13	Red Mtn.: Goldslide porphyry	197.6	1.9	1.9	4.4	4.4	40	40	U–Pb[g]	Rhys et al. 1995
14	Rupert Inlet[b]	191					52	52	U–Pb	Pálfy et al. 2000*b*
15	Ashman Ridge 1[b]	189					50	53	U–Pb	Pálfy et al. 2000*b*
16	Ashman Ridge 2[b]	192					50	53	U–Pb	Pálfy et al. 2000*b*
17	Telkwa Range 1	192.8	5.0	0.6	6.4	0.6	54	54	U–Pb[f]	Pálfy et al. 2000*b*
18	Telkwa Range 2	191.5	0.8	0.8	0.8	0.8	54	62	U–Pb[e]	Pálfy et al. 2000*b*
19	Joan Lake	193.1	2.1	3.7	4.8	3.7	50	62	U–Pb[h]	Thorkelson et al. 1995
20	Chuchi intrusion	188.5	2.5	2.5	2.5	2.5	62	64	U–Pb[e,i]	Nelson and Bellefontaine 1996
21	Atlin Lake (East shore)	187.5	1.0	1.0	1.0	1.0	62	62	U–Pb[e]	Mihalynuk and Gabites unpublished data, 1996
22	Todagin Mtn.	185.6	6.1	0.6	7.3	0.6	63	63	U–Pb[f]	Pálfy et al. 2000*b*
23	Atlin Lake (Sloko Island)[c]	186.6	0.5	1.0	0.5	1.0	50	64	U–Pb[i]	Johannson and McNicoll 1997
24	Atlin Lake (Copper Island)	185.8	0.7	0.7	0.7	0.7	64	64	U–Pb[e]	Johannson and McNicoll 1997
25	Skinhead Lake	184.7	0.9	0.9	0.9	0.9	64	64	U–Pb[e]	Pálfy et al. 2000*b*
26	Whitehorse	184.1	4.2	1.6	5.8	1.6	64	64	U–Pb[h]	Hart 1997
27	Eskay porphyry	184	5	1	6	1	65	82	U–Pb[f]	Macdonald et al. 1992; Childe 1996
28	McEwan Creek pluton	183.2	0.7	0.7	0.7	0.7	71	72	U–Pb[e]	Evenchick and McNicoll 1993
29	Mt. Brock range	180.4	10.5	0.4	11.2	0.4	71	72	U–Pb[f]	Pálfy et al. 2000*b*
30	Yakoun River	181.4	1.2	1.2	1.2	1.2	73	73	U–Pb[e]	Pálfy et al. 1997
31	Diagonal Mtn. 1	179.8	6.3	6.3	7.5	7.5	75	80	U–Pb[g]	Pálfy et al. 2000*b*
32	Julian Lake	178	1	1	1	1	75	75	U–Pb[e]	Mortensen and Lewis unpublished data, 1996
33	Treaty Ridge	177.3	0.8	0.8	0.8	0.8	82	92	U–Pb[e]	Friedman and Anderson unpublished data, 1997
34	Eskay rhyolite west	175.1	2.4	2.4	4.7	4.7	75	82	U–Pb[g]	Childe 1996
35	Eskay rhyolite east	174.1	2.1	1.1	4.5	1.1	75	82	U–Pb[h]	Childe 1996
36	Diagonal Mtn. 2	167.2	10.5	0.4	11.2	0.4	95	96	U–Pb[f]	Pálfy et al. 2000*b*
37	Gunlock	166.3	0.8	0.8	3.5	3.5	95	96	Ar–Ar	Kowallis et al. 1993
38	Burnaby Island Plutonic Suite	168	4	4	4	4	90	102	U–Pb[e]	Anderson and McNicoll 1995; Anderson and Reichenbach 1991
39	Harrison Lake	166	0.4	0.4	0.4	0.4	90	100	U–Pb[e]	Mahoney et al. 1995
40	McDonell Lake	158.1	2.0	1.5	2.0	1.5	102	102	U–Pb[e]	Pálfy et al. 2000*b*
41	Copper River	162.6	2.9	7.0	2.9	7.0	102	102	U–Pb[i]	Pálfy et al. 2000*b*
42	Diagonal Mtn. 3[b]	158.4	0.8	0.8	0.8	0.8	102	111	U–Pb[i]	Pálfy et al. 2000*b*
43	Chacay Melehué 1[b]	161	0.5	0.5	0.5	0.5	102	111	U–Pb[i]	Odin et al. 1992
44	Chacay Melehué 2	160.5	0.3	0.3	0.3	0.3	111	111	U–Pb[i,e]	Odin et al. 1992

Table 1 (*concluded*).

Item	Short name	Age (Ma)	Error Int. +2σ	Error Int. –2σ	Ext. +2σ	Ext. –2σ	Base[a]	Top[a]	Method	Reference
45	Tsatia Mtn.[c]	160.7	0.7	0.7	0.7	0.7	100	112	U–Pb[e]	Ricketts and Parrish 1992
46	Josephine ophiolite	162	7	2	8	2	40	120	U–Pb[f]	Harper et al. 1994
47	Rogue Formation[b]	157	3	3	3	3	90	120	U–Pb[e]	Harper et al. 1994; Saleeby 1984
48	Hotnarko volcanics	154.4	1.2	1.2	4.2	4.2	110	120	U–Pb[i]	van der Heyden 1991
49	Tidwell Member	154.9	1	1	3.5	3.5	130	130	Ar–Ar	Kowallis et al. 1998
50	Brushy Basin Member 1	150.3	0.5	0.5	3.0	3.0	130	130	Ar–Ar	Kowallis et al. 1998
51	Brushy Basin Member 2	150.2	1	1	3.5	3.5	130	130	Ar–Ar	Kowallis et al. 1998
52	Brushy Basin Member 3	149.3	1.1	1.1	3.6	3.6	130	140	Ar–Ar	Kowallis et al. 1998
53	Brushy Basin Member 4	149.3	1	1	3.5	3.5	130	140	Ar–Ar	Kowallis et al. 1998
54	Brushy Basin Member 5	149	0.8	0.8	3.3	3.3	130	140	Ar–Ar	Kowallis et al. 1998
55	Brushy Basin Member 6	148.1	1	1	3.5	3.5	130	140	Ar–Ar	Kowallis et al. 1998
56	Grindstone Creek	137.1	1.6	0.6	1.6	0.6	150	150	U–Pb[e]	Bralower et al. 1990

[a]Code numbers for constraining stratigraphic units (base and top) are given in Table 2.
[b]minimum age.
[c]maximum age.
[d]Weighted mean $^{206}Pb/^{238}U$ age
[e]Mean $^{206}Pb/^{238}U$ age with cumulative error
[f]$^{206}Pb/^{238}U$ age with asymmetric error (plus side from error of weighted mean $^{207}Pb/^{206}Pb$ age, minus side from error of $^{206}Pb/^{238}U$ age
[g]Weighted mean $^{207}Pb/^{206}Pb$ age
[h]Weighted mean $^{207}Pb/^{206}Pb$ age with asymmetric error (plus side from error of weighted mean $^{207}Pb/^{206}Pb$ age, minus side from error of $^{206}Pb/^{238}U$ age
[i]Lower intercept age.

hibit loss of radiogenic daughter products contribute only minimum ages for boundary estimation. Of the U–Pb ages, only data sets incorporating multiple fraction analyses are considered. Accepted zircon dates were obtained on abraded grains. The threshold of acceptable precision is generally ±5 Ma (unless stated otherwise, internal errors reported at 2 σ level). Less precise dates are only exceptionally included, where warranted by the lack of better data. Our maximum allowable error is lower than the 6 Ma (2σ) average precision in the data set of MTS94, the most recently produced Mesozoic time scale. Chronostratigraphic constraints are obtained using a variety of methods, but preference is given to ammonite biochronology. Assignment of age brackets is based on the tightest possible biochronologic constraints and they may be complemented with reasonable geologic age inferences. Isotopic ages are considered critical to the boundary estimation, if they are at least bracketed by adjacent stages. Through the use of zonal biochronology, the average chronostratigraphic precision is significantly improved, although exceptionally a few loosely constrained ages need to be included, where better dates are unavailable.

Table 1 lists the selected critical ages arranged by stages. Also included are several Late Triassic ages and one Early Cretaceous age that help constrain the lower and upper boundary of the Jurassic, bringing the total number of critical dates used to 56. A detailed discussion of the isotopic age and chronostratigraphic constraint of each database item is given in Appendix 1, complemented by Appendix 2 that consists of an annotated list of U–Pb and $^{40}Ar/^{39}Ar$ ages that were used in GTS89 and (or) MTS94 but are rejected in this study.[2]

Direct dating of stratigraphic boundaries

Ideally, the age of a biochronologically defined boundary is determined by isotopic dating of a volcanogenic layer situated at or in the immediate vicinity of the boundary. This situation occurs rarely in the Jurassic. The Triassic–Jurassic boundary is directly dated in both marine (Pálfy et al. 2000*a*) and continental sections (Dunning and Hodych 1990; Hodych and Dunning 1992). In the Queen Charlotte Islands, a tuff layer immediately below the boundary defined by radiolarian, conodont, and ammonoid biostratigraphy yielded a U–Pb age of 199.6 ± 0.3 Ma (item 5) (Pálfy et al. 2000*a*). In the Newark Basin, U–Pb ages of 200.9 ± 1.0 Ma (item 8) and 201.3 ± 1.0 Ma (item 7) (Dunning and Hodych 1990) were obtained on sills that are thought to be feeders to the Orange Mountain basalt, the lowermost extrusive rock which lies immediately above the Triassic–Jurassic boundary defined by palynology and vertebrate biostratigraphy (Fowell and Olsen 1993; Olsen et al. 1987). From the Fundy Basin, a U–Pb zircon age of $201.7^{+1.4}_{-1.1}$ Ma (Hodych and Dunning 1992) is reported from the North Mountain basalt, the base of which lies about 20 m above the Triassic–Jurassic boundary; because standard chronostratigraphy is based on marine biostratigraphy, we regard 199.6 ± 0.3 Ma as a close approximation of the beginning of Jurassic and infer a slight diachrony between marine and terrestrial extinctions (Pálfy et al. 2000*a*).

A volcanic ash layer directly above the base of the Middle Toarcian *Crassicosta* Zone, a regional standard ammonite zone for North America, is dated by U–Pb method at 181.4 ± 1.2 Ma in the Queen Charlotte Islands (Pálfy et al. 1997).

[2] Appendices 1 and 2 are available as supplemental material

A third directly dated level is the boundary of the South American *Steinmanni* and *Vergarensis* chrons that is equated to the Bathonian–Callovian boundary (Riccardi et al. 1991). A tuff layer located at this boundary in the Chacay Melehué section (Neuquén Basin, Argentina) yielded a U–Pb zircon date of 161.0 ± 0.5 Ma (Odin et al. 1992).

Chronogram estimation of boundaries

In the absence of direct isotopic dating, the ages of a stratigraphic boundary can be estimated using dates from adjacent units. The chronogram method, described in detail in GTS89 (Harland et al. 1990), calculates the error function value (E) for trial ages in a time window scanned for possible boundary ages. Taking into account the relevant isotopic dates, their stratigraphic position below or above the boundary in question, and their plus and minus errors, it provides a semi-rigorous measure of compatibility of the data with the trial ages. Following GTS89, the best chronogram estimate is defined by the minimum value of E or the mean of a range where E = 0, whereas the endpoints of the error range around the best estimates are taken where $E = E_{min} + 1$. We use here a slightly modified formula that accommodates asymmetric errors, which are common among interpreted U–Pb ages

$$\sum \frac{(Y_i - t_e)^2}{{S^{+}_{Yi}}^2} + \sum \frac{(O_i - t_e)^2}{{S^{-}_{Oi}}^2}$$

where t_e is the trial age, Y_i are the isotopic ages stratigraphically younger than the boundary for which $Y_i > t_e$, S^{+}_{Yi} are their plus 2σ external errors, O_i are the isotopic ages stratigraphically older than the boundary for which $O_i < t_e$, S^{-}_{Oi} are their minus 2σ external errors.

We calculated chronograms for each chron boundary in the Hettangian through Bajocian and each substage boundary in the Bathonian through Callovian. As not all isotopic dates have zonal or substage resolution constraints (not even attempted for the Oxfordian through Tithonian), stage boundary chronograms were also calculated (Table 2). These may be different from the chronogram of the earliest chron of the stage if significant dates lack zonal constraints (e.g., the Hettangian). The chronogram method assumes a random distribution of dates within the stratigraphic intervals whose boundaries are sought. Therefore, the directly determined boundary ages listed above are omitted from the calculation of those boundaries and accepted if they overlap with the error range of the respective chronogram.

If a unit lacks critical dates confined to it, its chronogram tends to converge to the next younger unit. Figure 2 thus shows only those meaningful chronograms that have no identical counterpart for a younger stratigraphic boundary.

The assumption of random distribution may prove to be unfounded for stage chronograms, if dates are crowded in some parts and are missing from other parts of the unit. This is demonstrated for the Sinemurian, Pliensbachian, Aalenian, Bajocian, and Bathonian, and its consequences are discussed next.

Adjusted stage boundary estimates

Stage boundaries are of particular interest, therefore, their age estimates are considered individually. Chronogram estimates may be biased if sufficient data do not exist in the vicinity of the boundary. This is indicated if the chronogram of the earliest chron is identical to any or both adjacent chrons. While chronogram maxima and minima are retained as valid, reasonable adjustments in the best estimates are made with respect to the critical data.

As discussed above, two stage boundaries are dated directly. The base of the Hettangian, fixed at 199.6 ± 0.4 Ma based on the directly dated marine boundary, is in good agreement with the chronogram age derived from the other pertinent dates. The base of the Callovian is pegged by a single date (161.0 ± 0.5 Ma, item 43) that is consistent with a slightly younger date (item 44) produced by the same authors (Odin et al. 1992) from the same section, but is in conflict with another relatively precise age (item 40) from the Bathonian. As a result, the chronogram minimum takes a value of ca. 2.5 and, in this case, we choose to define the error range where $E = 2E_{min}$. Such a chronogram age of $160.4^{+1.1}_{-0.5}$ Ma overlaps with the directly determined boundary age and is our preferred estimate.

In the Early Jurassic, the chronogram age of the Pliensbachian–Toarcian boundary ($183.6^{+1.7}_{-1.1}$ Ma) is tightly controlled, as there is a series of good quality chronograms for the neighbouring chrons. The Hettangian–Sinemurian and the Sinemurian–Pliensbachian boundaries are less well constrained. The Canadensis Zone, likely to span the Hettangian–Sinemurian boundary, is here arbitrarily classified as Sinemurian, but this does not affect the chronogram. Early and early Late Sinemurian ages are all minimum ages causing an asymmetric chronogram. The chronogram estimate of $195.4^{+2.8}_{-4.6}$ Ma is adjusted by shifting the best estimate up to $196.5^{+1.7}_{-5.7}$ Ma. Such reapportioning is also supported by observed sediment thicknesses in many Lower Jurassic sections.

The chronogram of the Sinemurian–Pliensbachian is identical to the initial Whiteavesi chron boundary and is not well constrained. On the other hand, the chronogram of the Late Sinemurian Harbledownense chron does not differ from that of the initial Sinemurian boundary. No data exist from the *Tetraspidoceras* and *Imlayi* chrons adjacent to the Sinemurian–Pliensbachian boundary. A possible less conservative interpretation of item 17, the main control on the Late Sinemurian, would also suggest a younger age (192–194 Ma) for the *Harbledownense* chron. Therefore we pick 191.5 Ma (with errors adjusted to $^{+1.9}_{-4.7}$ Ma) for the best estimate of the age of the Sinemurian–Pliensbachian boundary.

In the Middle Jurassic, the tight Toarcian–Aalenian boundary chronogram is identical to the middle–late Aalenian boundary chronogram calling for an upward shift in age. We propose a boundary age of 178.0 Ma (with errors adjusted to $^{+1.0}_{-1.5}$ Ma).

The Aalenian–Bajocian and Bajocian–Bathonian boundary chronograms are less tightly constrained and are also controlled by dates from the younger half of the stages. Their best estimates and maximum ages are only marginally older than those of the Late Bajocian Rotundum Zone and the Middle–Late Bathonian, respectively. Hence, an upward adjustment of boundary ages is likely to produce more realistic allocation. We propose 174 Ma for the initial Bajocian boundary, corresponding to the oldest trial age, with an error function value of zero. The duration of the late Aalenian

Table 2. Chronogram estimates of the initial boundary of chrons, substages, and stages.

Chron, substage, stage	Code	Initial boundary	Maximum	Minimum	+Error (2σ)	–Error (2σ)	Error range
Psiloceras–Euphyllites	41	201.1	202.3	197.2	1.2	3.9	5.1
Franziceras	**42**	**199.6**	**200.0**	**197.2**	**0.4**	**2.4**	**2.8**
Pseudoaetomoceras	43	196.6	202.2	190.7	5.6	5.9	11.5
Canadensis	51	195.3	198.2	190.7	2.9	4.6	7.5
Coroniceras–Arnouldi	52	195.3	198.2	190.7	2.9	4.6	7.5
Varians	53	195.3	198.2	190.7	2.9	4.6	7.5
Harbledownense	54	195.3	198.2	190.7	2.9	4.6	7.5
Tetraspidoceras	55	190.7	193.4	186.8	2.7	3.9	6.6
Imlayi	61	190.7	193.4	186.8	2.7	3.9	6.6
Whiteavesi	62	190.7	193.4	186.8	2.7	3.9	6.6
Freboldi	**63**	**186.7**	**188.5**	**185.1**	**1.8**	**1.6**	**3.4**
Kunae	**64**	**185.7**	**186.2**	**185.1**	**0.5**	**0.6**	**1.1**
Carlottense	**65**	**184.1**	**185.3**	**182.5**	**1.2**	**1.6**	**2.8**
Kanense	**71**	**183.7**	**185.3**	**182.5**	**1.6**	**1.2**	**2.8**
Planulata	72	182	185.3	177.1	3.3	4.9	8.2
Crassicosta	73	180.1	180.8	177.1	0.7	3.0	3.7
Hillebrandti	74	180.1	180.8	177.1	0.7	3.0	3.7
Yakounensis	**75**	**180.1**	**180.8**	**177.1**	**0.7**	**3.0**	**3.7**
Westermanni–Scissum	81	177.6	179.0	176.5	1.4	1.1	2.5
Howelli	**82**	**177.6**	**179.0**	**176.5**	**1.4**	**1.1**	**2.5**
Widebayense	91	170.7	175.2	166.0	4.5	4.7	9.2
Crassicostatus	92	170.7	175.2	166.0	4.5	4.7	9.2
Kirschneri	93	170.7	175.2	166.0	4.5	4.7	9.2
Oblatum	94	170.7	175.2	166.0	4.5	4.7	9.2
Rotundum	95	170.7	175.2	166.0	4.5	4.7	9.2
Epizigzagiceras	96	168.4	175.2	160.3	6.8	8.1	14.9
Middle–Late Bathonian	102	164.5	169.8	160.3	5.3	4.2	9.5
Early Callovian	**111**	**160.4**	**161.5**	**159.9**	**1.1**	**0.5**	**1.6**
Middle Callovian	112	156.5	159.6	151.4	3.1	5.1	8.2
Late Callovian	113	156.5	159.6	151.4	3.1	5.1	8.2
HETTANGIAN	**40**	**201.1**	**202.3**	**197.5**	**1.2**	**3.6**	**4.8**
SINEMURIAN	50	195.4	198.2	190.8	2.8	4.6	7.4
PLIENSBACHIAN	60	190.7	193.4	186.8	2.7	3.9	6.6
TOARCIAN	**70**	**183.6**	**185.3**	**182.5**	**1.7**	**1.1**	**2.8**
AALENIAN	**80**	**177.6**	**179.0**	**176.5**	**1.4**	**1.1**	**2.5**
BAJOCIAN	90	171	175.2	166.1	4.2	4.9	9.1
BATHONIAN	100	164.5	169.8	160.4	5.3	4.1	9.4
CALLOVIAN	**110**	**160.4**	**161.5**	**159.9**	**1.1**	**0.5**	**1.6**
OXFORDIAN	120	156.5	159.6	151.4	3.1	5.1	8.2
KIMMERIDGIAN	130	154.7	158.5	151.4	3.8	3.3	7.1
TITHONIAN	140	145.8	158.5	135.5	12.7	10.3	23.0
BERRIASIAN	150	145.8	158.5	135.5	12.7	10.3	23

Note: Chrons with poor stratigraphic record (Euphyllites, Coroniceras, and Westermanni) were combined with adjacent chrons. Bold face designates good quality chronograms that are different from the next younger stratigraphic boundary, and the error range is less than 5 Ma.

Howelli chron thus appears to be 3.6 Ma, conspicuously longer than any other chron. There are three controlling dates from the Howelli Zone (items 33–35) and none from the adjacent zones, therefore, the unusually long chron duration may be an artifact of a slight inaccuracy in some of the isotopic dates. The preferred estimate for the Bajocian–Bathonian boundary is 166 Ma.

The chronogram boundary estimates suggest an extremely short duration for the Oxfordian stage, which appears unrealistic. This weakness derives from the sparseness of Late Jurassic isotopic ages. Instead of arbitrary adjustments, we postpone its resolution until further data become available.

Latest Jurassic stage boundary estimates through magnetochronologic interpolation

The lack of isotopic ages from definitively dated Tithonian rocks renders chronogram estimation of the Kimmeridgian–Tithonian and the Tithonian–Berriasian boundaries extremely imprecise. Consequently, we employ

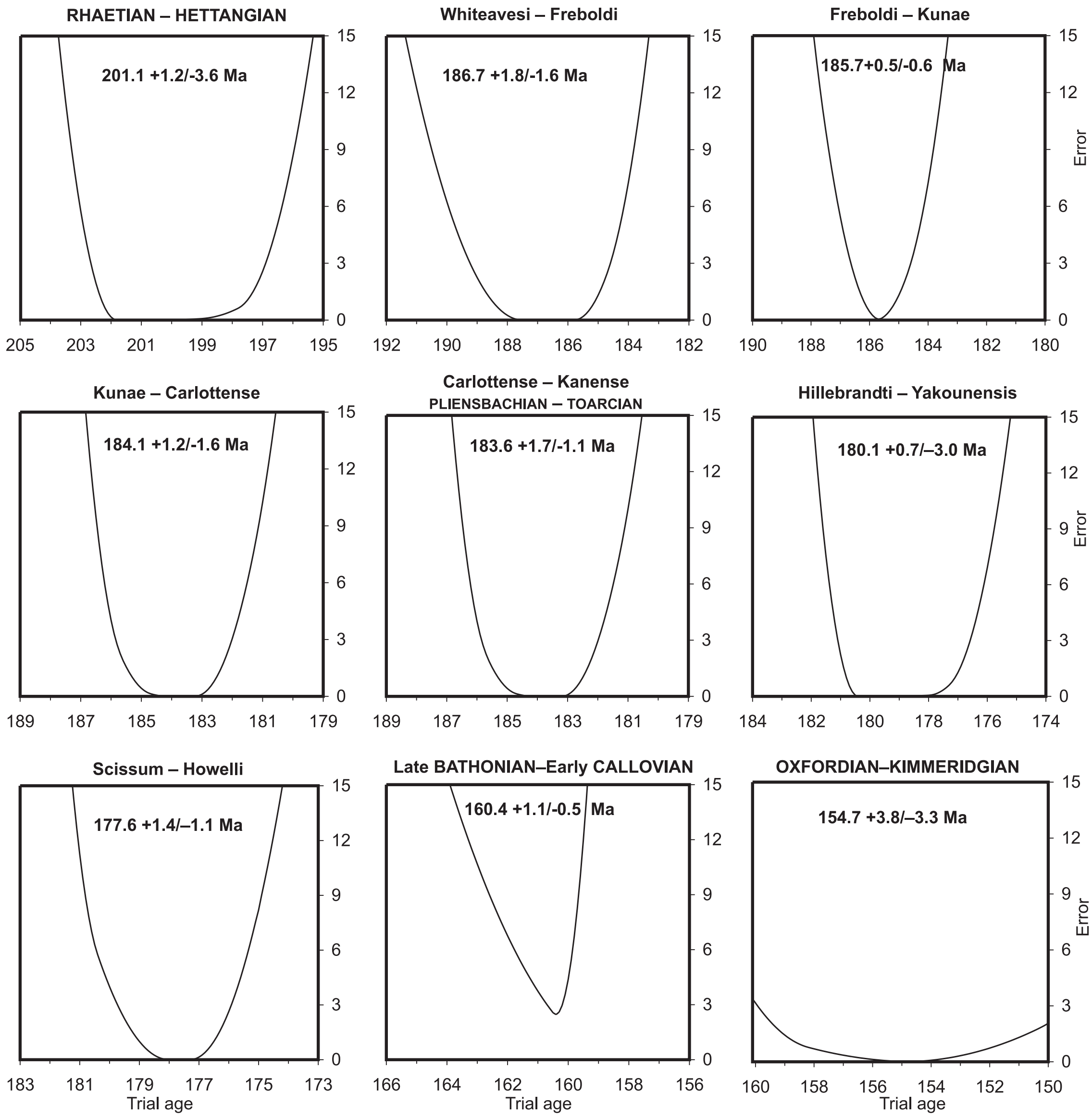

Fig. 2. Chronogram plots of the best constrained chronostratigraphic boundary ages. Scale is uniform to facilitate comparison, showing a 10 Ma time window for trial ages and a 0–15 scale on the error value axis for chronograms with a less than 5 Ma range of error. The lower quality Oxfordian–Kimmeridgian boundary chronogram is also shown because of its importance as a tie-point for interpolation in the Late Jurassic.

magnetochronologic interpolation similar to that discussed in detail in GTS89 and MTS94. For the Jurassic–Cretaceous transition, we use the magnetic anomaly block model derived from the Hawaiian lineation set that was shown to best approximate a constant spreading rate (Channell et al. 1995). The oldest Cretaceous isotopic date available to anchor the magnetochronology is $137.1^{+1.6}_{-0.6}$ Ma (item 56) that is indirectly correlated to the Late Berriasian M16 chron using nannofossils (Bralower et al. 1990). A subsequent revision suggests that a broader correlation with the M16–M15 interval is more appropriate (Channell et al. 1995). Anchoring the Jurassic side of the lineation set is more controversial. MTS94 uses direct dating of M26r at 155.3 ± 3.4 Ma in the Argo Abyssal Plain (Ludden 1992). This, in fact, is a minimum K–Ar age of a celadonite vein, therefore, we did not include it in our data set. From the same site, incremental heating $^{40}Ar/^{39}Ar$ dating of basalt from M25–26 did not yield a plateau age, only a disputable total fusion age of 155 ± 6 Ma (2σ) was obtained (Ludden 1992).

Instead we rely on the chronogram estimate of the Oxfordian–Kimmeridgian boundary ($154.7^{+3.8}_{-3.3}$ Ma). Magnetochronologic correlation of land-based, ammonite-constrained sections with the oceanic magnetic anomalies allow the placement of this boundary at the base of M25n (Ogg and Gutowski 1996). As the isotopic ages controlling the chronogram are not precisely constrained (items 48–55), allowance needs to be made for correlation uncertainty in the chronogram age. We regard M29 near the base of the Late Oxfordian (Ogg and Gutowski 1996) as a maximum, based on the maximum age of the Morrison Formation by ostracods and charophytes (Schudack et al. 1998) and ammonoids from underlying strata (Callomon 1984; Imlay 1980). The interpolation yields $150.5^{+3.4}_{-2.8}$ Ma for the Kimmeridgian–Tithonian boundary and $141.8^{+2.5}_{-1.8}$ Ma for the Tithonian–Berriasian boundary. The error limits reflect a combination of the numeric errors of the Oxfordian–Kimmeridgian boundary chronogram and the Berriasian isotopic age and the associated biochronologic–magnetochronologic correlation uncertainties.

The Jurassic time scale

In summary, the initial boundaries of Jurassic stages are proposed as follows in Table 3.

Table 3. The Jurassic time scale.

Stage	Age
Berriasian (Jurassic–Cretaceous)	$141.8^{+2.5}_{-1.8}$ Ma
Tithonian	$150.5^{+3.4}_{-2.8}$ Ma
Kimmeridgian	$154.7^{+3.8}_{-3.3}$ Ma
Oxfordian	$156.5^{+3.1}_{-5.1}$ Ma
Callovian	$160.4^{+1.1}_{-0.5}$ Ma
Bathonian	$166.0^{+3.8}_{-5.6}$ Ma
Bajocian	$174.0^{+1.2}_{-7.9}$ Ma
Aalenian	$178.0^{+1.0}_{-1.5}$ Ma
Toarcian	$183.6^{+1.7}_{-1.1}$ Ma
Pliensbachian	$191.5^{+1.9}_{-4.7}$ Ma
Sinemurian	$196.5^{+1.7}_{-5.7}$ Ma
Hettangian (Triassic–Jurassic)	199.6 ± 0.4 Ma

In addition, Early and Middle Jurassic chron boundary ages were also estimated. Although this list remains incomplete, the following initial chron boundaries (other than those coinciding with stage boundaries) are known with less than 5 Ma range of error: Howelli (Aalenian), $177.6^{+1.4}_{-1.1}$ Ma; Yakounensis (Toarcian), $180.1^{+0.7}_{-3.0}$ Ma; Crassicosta (Toarcian), 181.4 ± 1.2 Ma; Carlottense (Pliensbachian), $184.1^{+1.2}_{-1.6}$ Ma; Kunae (Pliensbachian), $185.7^{+0.5}_{-0.6}$ Ma; Freboldi (Pliensbachian), $186.7^{+1.8}_{-1.6}$ Ma.

Discussion

The Jurassic time scale developed here differs from its predecessors in several aspects of methodology. It is based exclusively on U–Pb and $^{40}Ar/^{39}Ar$ ages. Although it was not an explicit criterion for inclusion in the isotopic database, all ages used were reported after 1990. Despite such restrictions, there are 50 Jurassic isotopic ages used in this study, slightly exceeding the similar number in previous scales (e.g., 45 in GTS89 and 43 in MTS94). Both GTS89 and MTS94 entered the errors at the 1σ confidence level, whereas the more conservative 2σ level is used throughout this study. Validity of comparison between U–Pb and $^{40}Ar/^{39}Ar$ ages was ensured by the use of external errors that reflect decay constant uncertainty.

Improvement is most significant in the Early Jurassic portion of the time scale. Through direct estimates of the base of the Hettangian and the Crassicosta chron and five additional, precise (i.e., error range is less than 5 Ma) chronogram boundary ages, one-third of the 20 chrons now have acceptable boundary estimates. Direct estimation of zonal duration was attempted for parts of the Pliensbachian and Toarcian only. From the Freboldi to the Crassicosta chrons, a sequence of five well constrained chronograms and a directly dated boundary permits us to propose the following best estimates for zonal duration: Freboldi: 1.0 Ma, Kunae: 1.6 Ma, Carlottense: 0.4 Ma, Kanense: 1.7 Ma, and Planulata: 0.6 Ma. Although the errors of boundary ages clearly indicate that chron durations are also subject to error, the observed fourfold difference among the chron durations is unlikely to be accounted for by random error. Therefore, Jurassic ammonite chrons appear to be of disparate length. Such a conclusion seriously undermines earlier time scales that were founded on interpolation based on the assumption of equal duration of ammonite chrons.

Figure 3 presents a comparison of the proposed time scale with the five scales most widely used in recent works. Key observations are as follows: (1) the Jurassic Period had a duration of approximately 58 Ma, shorter than previously estimated; (2) the Triassic–Jurassic boundary (~200 Ma) is younger than previously thought; (3) the proposed age of the Jurassic–Cretaceous boundary (142 Ma) is close to that suggested by the DNAG, GTS89, and MTS94 scales, whereas the same boundary is too young in EXX88 and OD94, which use K–Ar glauconite ages; (4) the Late Jurassic appears to be shorter (15 Ma) than the Early and Middle Jurassic epochs (22 and 21 Ma, respectively); (5) reapportioning of time in the Early and Middle Jurassic is a consequence of not using interpolated stage durations.

As demonstrated here, the new generation of time scales needs to be based exclusively on U–Pb and $^{40}Ar/^{39}Ar$ ages to benefit from the improved precision and accuracy offered by these two dating methods. Improved statistical techniques can be sought to refine age estimates while accommodating asymmetric, non-Gaussian errors frequently encountered in U–Pb dating. However, the goal of systematically defining chron boundary ages can only be achieved through the acquisition of still more calibration points, ultimately eliminating the need for interpolation. The dating of volcanic flow or pyroclastic units within fossiliferous sequences is proved to be the most successful way of obtaining calibration points. An increasingly more refined time scale offering consistently high precision and resolution at the zonal level is shown in this study to be a realistic goal.

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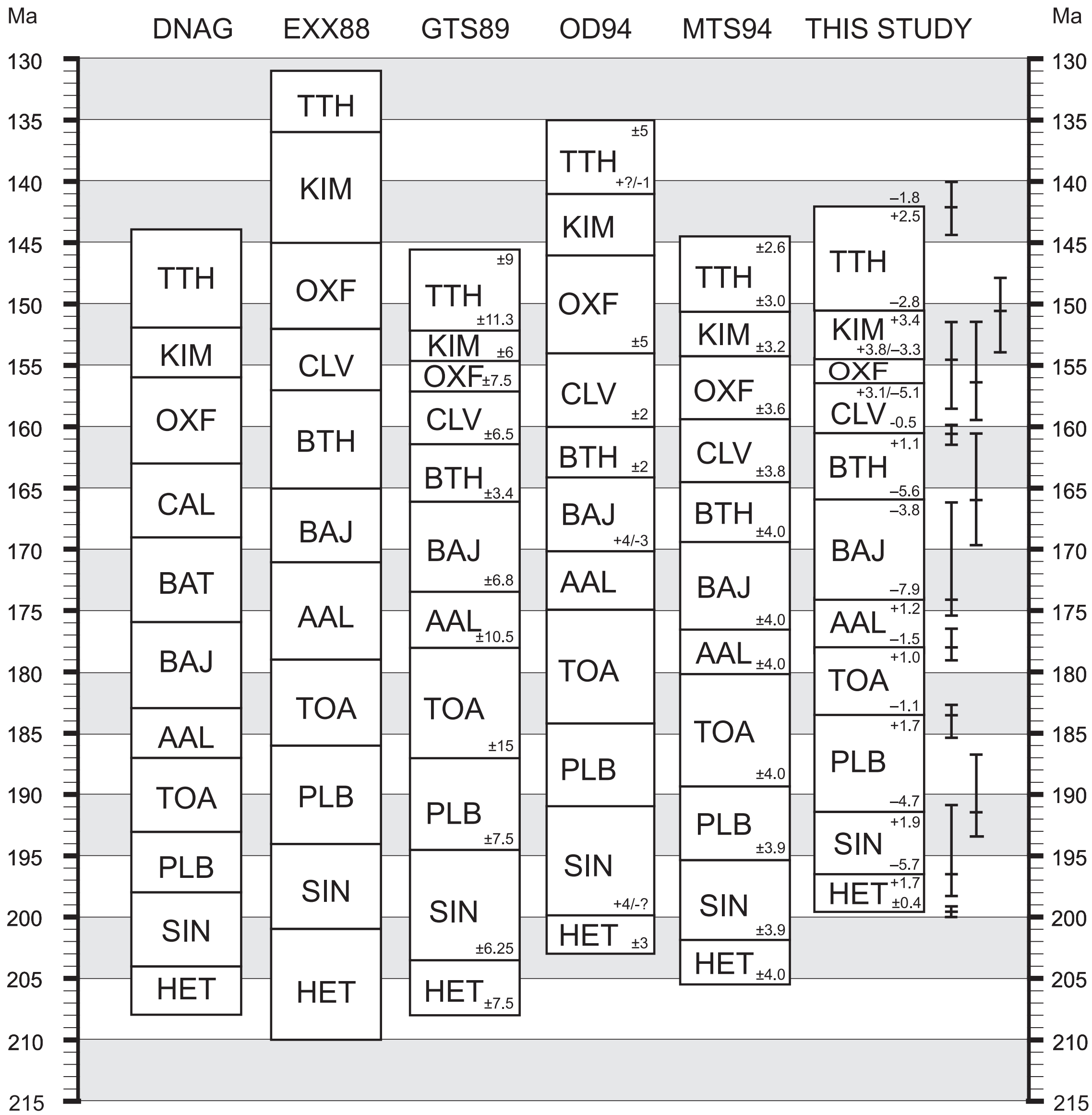
Fig. 3. Comparison of the proposed Jurassic time scale with other major time scales (DNAG, EXX88, GTS89, MTS94, OD94). Stage abbreviations follow those in Harland et al. (1990). Error bars for stage boundaries in the new scale are shown on the right.



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Appendix 1.

Comments on items used in the isotopic database

A critical evaluation is given for the isotopic ages and their stratigraphic constraints used for the revised Jurassic time scale. Only brief reference is made to dates published recently by the team of present authors (Pálfy et al. 1997, 1999, 2000*a*, 2000*b*). Errors quoted here are internal errors at 2σ level, although for chronogram calculations we used external errors (see Table 1).

Item 1 — Guichon Creek batholith

A U–Pb zircon age of 210 ± 3 Ma was obtained (Mortimer et al. 1990) and interpreted as the crystallization age of the Guichon Creek batholith. The dated sample is not closely associated with fossiliferous strata but sedimentary intercalations within the Nicola Group that are crosscut by the batholith range up to the Middle Norian (Frebold and Tipper 1969). A minimum age for the intrusion is no older than Pliensbachian based on ammonites from the onlapping Ashcroft Formation. K–Ar and Rb–Sr minimum ages averaged around 205–209 Ma from the Guichon Creek batholith appear as NDS177 (Armstrong in Odin 1982). The dates are subsequently used as 205 ± 2.5 Ma (1σ) with Norian–Hetangian brackets in GTS89. MTS94 cites the same data without considering the more precise and accurate U–Pb age used here.

Items 2–3 — Griffith Creek volcanics

Two U–Pb ages (205.8 ± 0.9 and $205.8^{+1.5}_{-3.1}$ Ma) were obtained from the Griffith Creek volcanics in northwest British Columbia (Thorkelson et al. 1995), a volcanic unit that is tightly folded and overlies a conglomerate containing Upper Triassic limestone pebbles and Norian clastic strata of the Stuhini Group. Folding predated deposition of the younger, Lower Jurassic Cold Fish volcanics and records the regional deformational event near the Triassic–Jurassic boundary (Thorkelson et al. 1995).

Items 4 and 13 — Goldslide intrusions, Red Mountain
Item 12 — Cambria Icefield

The Biotite porphyry, the oldest of three phases of the mineralized Goldslide intrusions on Red Mountain near Stewart (Rhys et al. 1995), yielded a 201.8 ± 0.5 Ma U–Pb date on zircon (Greig et al. 1995). Only this older phase is affected by pervasive cleavage related to a regional deformation event near the Triassic–Jurassic boundary (Greig et al. 1995). A few tens of kilometres to the north, the youngest deformed strata contain the Rhaetian ammonoid *Choristoceras* (Jakobs and Pálfy 1994). Triassic radiolaria ranging in age from Ladinian–Carnian to Norian were recovered from strata cut by the Biotite porphyry (F. Cordey, personal communication 1996). The Goldslide porphyry, the youngest postkinematic phase was U–Pb zircon dated at 197.6 ± 1.9 Ma (Rhys et al. 1995). A tuff 8 km to the south that is considered correlative to the top of the Red Mountain succession yielded a U–Pb age of 199 ± 2 Ma (Greig and Gehrels, 1995). On Red Mountain, abundant peperitic structures and strongly disrupted country rocks with evidence of soft-sediment deformation suggest that at least some of the Goldslide intrusions intermingled with unlithified, wet sediments and are nearly coeval with their country rocks (Greig and Gehrels 1995; Rhys et al. 1995). We recovered an indeterminate ammonite of Early Jurassic aspect along with the bivalve *Oxytoma* stratigraphically above the synsedimentary intrusions. In Europe, *Oxytoma* is known as low as the Upper Triassic, but in the Eastern Pacific it first appears in the Lower Jurassic with common occurrences in the Hettangian (Aberhan 1994).

Item 5 — Kunga Island

A tuff layer from the marine sedimentary Sandilands Formation yielded a U–Pb zircon age of 199.6 ± 0.3 Ma (Pálfy et al. 2000*a*). The dated layer lies near the top of the uppermost Triassic *Globolaxtorum tozeri* radiolarian zone, immediately below the Triassic–Jurassic boundary as defined by radiolarian, conodont and ammonoid biochronology.

Items 6–8 — Newark Supergroup basalts

Three U–Pb dates exist for mafic, rift-related volcanic, and hypabyssal rocks of eastern North America. A zircon and baddeleyite age of 200.9 ± 1.0 Ma was obtained from the Palisades sill and a zircon age of 201.3 ± 1.0 Ma from the Gettysburg sill in the Newark Basin (Dunning and Hodych 1990). The sills are thought to be feeders to the Orange Mountain basalt, the lowermost extrusive rocks which immediately overlies the purported Triassic–Jurassic boundary based on palynological and vertebrate faunal evidence (Olsen et al. 1987; Fowell and Olsen 1993). From the Fundy Basin, a U–Pb zircon age of $201.7^{+1.4}_{-1.1}$ Ma (Hodych and Dunning 1992) is reported from the North Mountain basalt whose base lies some 20 m above the Triassic–Jurassic boundary, also defined by palynology and vertebrate biostratigraphy (Olsen et al. 1987).

Items 9–11 — Puale Bay

Three U–Pb zircon dates were obtained from Puale Bay, Alaska Peninsula, for samples whose age is constrained by ammonite biochronology at the zonal level (Pálfy et al. 1999). A Middle Hettangian (Liasicus Zone equivalent) tuff layer from near the top of the Kamishak Formation is dated at $200.8^{+2.7}_{-2.8}$ Ma. Tuffs from the overlying Talkeetna Formation are bracketed by Middle and Late Hettangian ammonites and yield crystallization ages of $197.8^{+1.2}_{-0.4}$ Ma and 197.8 ± 1.0 Ma.

Items 12 and 13 — see item 4
Item 14 — Rupert Inlet

Tuff layers occur in the Sinemurian Harbledown Formation exposed near Rupert Inlet on northern Vancouver Island. Ammonoids from below and above the sampled tuff indicate the late Early Sinemurian Arnouldi Assemblage. A tuff sample yielding few zircons gave only a minimum age of 191.3 ± 0.4 Ma (Pálfy et al. 2000*b*), based on a discordant fraction with the oldest $^{206}Pb/^{238}U$ age.

Items 15–16 — Ashman Ridge

Two U–Pb zircon ages were obtained from the Telkwa Formation on Ashman Ridge, British Columbia (Pálfy et al. 2000*b*). A densely welded rhyolite tuff yielded a minimum age of 189 Ma. Poor zircon recovery precluded a more definitive age determination. The minimum age is based on a near-concordant fraction with the oldest $^{206}Pb/^{238}U$ age, whereas other fractions indicate Pb loss. The dated sample was collected more than 200 m below a locally very fossiliferous, calcareous tuffaceous sandstone. The ammonoid fauna belongs to the early Late Sinemurian Varians Assemblage (Pálfy and Schmidt 1994).

Another sample from an andesite flow that immediately underlies the fossiliferous sedimentary rocks provided a minimum U–Pb zircon age of 192 Ma. The interpreted age is based on a concordant fraction with the oldest $^{206}Pb/^{238}U$ age, whereas iother fractions exhibit effects of Pb loss and (or) inheritance. The Telkwa Formation appears to represent geologically rapid accumulation of predominantly subaerially deposited volcanic and volcaniclastic rocks (Tipper and Richards 1976), hence the age of the isotopically dated volcanic units is considered to be early Late Sinemurian or only slightly older.

Items 17–18 — Telkwa Range

Two U–Pb zircon ages were obtained from a section of volcanic, volcaniclastic, and carbonate rocks of the Telkwa Formation in the southern Telkwa Range, British Columbia (Pálfy et al. 2000*b*). A dacite tuff unit was dated at $194.0^{+9.1}_{-1.8}$ Ma. *Paltechioceras* cf. *boehmi* and other ammonoids characteristic to the Upper Sinemurian Harbledownense Assemblage occur both below and above the tuff (Pálfy and Schmidt 1994). A rhyolite flow some 100 m farther upsection yielded an age of 191.5 ± 0.8 Ma.

Despite the stratigraphic distance, no significant difference in age is implied as rapid deposition of thick volcanogenic strata is typical in the proximal volcanic facies of the Hazelton Group (Tipper and Richards 1976). Faunas younger than Late Sinemurian are not known from the Telkwa Formation. Regionally, the oldest faunas reported from the overlying Nilkitkwa Formation are of Early Pliensbachian (Whiteavesi Zone) age (Tipper and Richards 1976).

Item 19 — Joan Lake

A U–Pb zircon age of $193.9^{+1.1}_{-4.5}$ Ma was reported from the Joan Lake area (northwestern British Columbia) (Thorkelson et al. 1995). The age interpretation is based on the weighted mean $^{207}Pb/^{206}Pb$ age of three of the four analyzed sections. The lower error limit is taken from the $^{206}Pb/^{238}U$ age and error of the oldest concordant fraction, whereas the upper error limit is that of the weighted mean $^{207}Pb/^{206}Pb$ age. Using all four fractions, we recalculated the age to $193.1^{+2.1}_{-3.7}$ Ma. The sample was collected near the top of a thick welded rhyolite tuff unit, which is disconformably overlain by the fossiliferous Joan Formation that contains an abundant ammonite fauna of Early Pliensbachian (Whiteavesi Zone) age (Thomson and Smith 1992). A time gap of undetermined duration is suggested by the erosional surface and the basal conglomerate at the contact between the two formations. The biochronological data, therefore, provide an upper bracket to the isotopic age: the Cold Fish volcanics at Joan Lake is Whiteavesi Zone (Early Pliensbachian) or older.

Three other samples from the Cold Fish volcanics in the Spatsizi River map area were also dated (Thorkelson et al. 1995). Rhyolite sills and a dyke yielded U–Pb ages of 193.7 ± 1.9, $194.8^{+1.0}_{-3.0}$, and 196.6 ± 1.6 Ma, respectively, all within error with the Joan Lake rhyolite tuff.

In MTS94, (Gradstein et al. 1994) (item 276) use a weighted mean of three of these ages arbitrarily simplified to symmetric errors and interpret it as Early Pliensbachian. Although there are limited occurrences of fossiliferous Lower Pliensbachian sediments interbedded with volcanic rocks elsewhere in the map area (Thomson et al. 1986), the evidence from the Joan Lake section itself doesn't justify this restrictive interpretation.

Item 20 — Chuchi intrusion

In the Mt. Milligan map area in northern Quesnellia, British Columbia, the locally fossiliferous Chuchi Lake Formation of the Takla Group is intruded by small igneous bodies. One of them, a monzonite intrusion near the BP–Chuchi property, was dated as 188.5 ± 2.5 Ma by U–Pb method (Nelson and Bellefontaine 1996). The crystallization age is defined by two concordant and overlapping titanite fractions and the lower intercept of a discordia line through two zircon fractions containing small amounts of inherited Proterozoic Pb component. Field observations (e.g., wet sediment deformation near the intrusive contacts, predominance of sills, crosscutting relationships) suggest synsedimentary (i.e., prelithification) emplacement of this high-level intrusion (Nelson and Bellefontaine 1996). Sedimentary beds in the area, likely correlative with the sedimentary rocks intruded by the dated monzonite body (Nelson and Bellefontaine 1996), yielded ammonites (*Leptaleoceras*, *Arieticeras*, and *Amaltheus*) of the Late Pliensbachian Kunae Zone (Tipper in Nelson and Bellefontaine 1996). In one section, however, an Early Pliensbachian (Whiteavesi Zone) ammonite fauna was recovered from a thin and stratigraphically lower sedimentary unit (Tipper 1996). A conservative interpretation is to bracket the crystallization age between the *Whiteavesi* and *Kunae* zones.

Items 21, 23, and 24 — Atlin Lake

The fossiliferous Lower Jurassic Laberge Group is well exposed on the shores of Atlin Lake in northwestern British Columbia and contains minor volcanic rocks (Johannson et al. 1997). A tuff layer (informally assigned to the Nordenskiold volcanics) yielded a U–Pb zircon date of 187.5 ± 1.0 Ma (item 21, M.G. Mihalynuk and J.E. Gabites, personal communication 1996). Ammonite collections stratigraphically below (*Tropidoceras actaeon*) and above (*Metaderoceras* cf. *talkeetnaense*, *Dubariceras* cf. *silviesi*, *Acanthopleuroceras* cf. *thomsoni*) the tuff indicate the Whiteavesi Zone (Early Pliensbachian) (Johannson 1994). On Copper Island, another crystal tuff layer stratigraphically higher was U–Pb dated at 185.8 ± 0.7 Ma (item 24) (Johannson and McNicoll 1997). This sample is also tightly constrained by ammonite biochronology. Diagnostic taxa, indicative of the Late Pliensbachian Kunae Zone, include *Reynesoceras colubriforme* and *Arieticeras* from below and *Leptaleoceras*, *Arieticeras*, and *Protogrammoceras* from above the tuff (Johannson 1994).

Another U–Pb age of 186.6 +0.5/–1.0 Ma was obtained from a granitoid boulder within a conglomerate on Sloko Island (item 23) (Johannson and McNicoll 1997). The conglomerate is also of Kunae Zone age based on the ammonite fauna (*Reynesoceras, Leptaleoceras, Protogrammoceras, Fuciniceras, Arieticeras*) recovered from adjacent finer-grained sedimentary rocks (Johannson 1994). Assuming geologically rapid uplift and erosion, the U–Pb date provides a useful maximum age for the Kunae Zone (Johannson et al. 1997).

Item 22 — Todagin Mountain

Lower Jurassic volcanic, volcaniclastic and sedimentary rocks of the Hazelton Group are exposed southwest of Todagin Mtn., northwestern British Columbia (Ash et al. 1996, 1997). A waterlain lapilli tuff, under- and overlain by shale, siltstone, and sandstone of *Freboldi* Zone (Early Pliensbachian) age, proven by the zonal index *Dubariceras freboldi*, was dated by U–Pb zircon method as $185.6^{+6.1}_{-0.6}$ Ma (Pálfy et al. 2000*b*). Another felsic volcanic unit exposed nearby yielded a preliminary U–Pb age of $181.0^{+5.9}_{-0.4}$ Ma (Ash et al. 1997). This unit is separated from the measured section by covered areas, rendering its stratigraphic relationships somewhat uncertain. Therefore, this date is not used for time scale calibration.

Item 25 — Skinhead Lake

A U–Pb age of 184.7 ± 0.9 Ma was obtained by M. Villeneuve from zircon in a rhyolite tuff near Skinhead Lake (west of Babine Lake, northwestern British Columbia) (Pálfy et al. 2000*b*). A volcanogenic sandstone intercalation above the dated volcanic unit yielded ammonoids (*Arieticeras* sp., and *Fanninoceras* sp.) of the Upper Pliensbachian Kunae Zone. As similar ammonoid faunas are known from dark shale underlying the volcanic package in the Fulton Lake area (Tipper and Richards 1976; H.W. Tipper, personal communication, 1996), the U–Pb dated unit is regarded of Kunae Zone age.

Item 26 — Whitehorse

A dacite tuff assigned to the Nordenskiold volcanics is U–Pb dated at $184.1^{+4.2}_{-1.6}$ Ma near Whitehorse, Yukon (Hart 1997). It is interbedded in a section of fossiliferous Laberge Group sediments. *Arieticeras* occurs a few metres below the tuff whereas *Arieticeras* and *Amaltheus* cf. *stokesi* was collected upsection indicating the presence of the Upper Pliensbachian Kunae Zone. The section is discussed in detail along with illustration of the ammonite fauna by Pálfy and Hart (1995).

Item 27 — Eskay porphyry

A sill-like feldspar porphyry intrusion in near the Eskay Creek gold mine in the Iskut River area (northwestern British Columbia) was U–Pb zircon dated as 186 ± 2 Ma (Macdonald et al. 1992). It was recalculated and reinterpreted as 184^{+5}_{-1} Ma (Childe 1996). The porphyry intrudes fossiliferous mudstone that yielded *Lioceratoides propinquum*, *Protogrammoceras* cf. *kurrianum* and other hildoceratids characteristic to the Pliensbachian–Toarcian transition (Nadaraju 1993). The ammonoids range from the topmost Pliensbachian Carlottense Zone to the basal Toarcian Kanense Zone, although the absence of *Dactylioceras* favours an assignment to the Carlottense Zone, for which the age of the intrusion is regarded as a minimum age.

Item 28 — McEwan Creek pluton

The quartz monzonite McEwan Creek pluton intrudes Lower Jurassic and older strata in the northwest part of the Spatsizi River map area (northwest British Columbia) and was dated by U–Pb method (Evenchick and McNicoll 1993). The reported age of 183.5 ± 0.5 Ma is based on the $^{206}Pb/^{238}U$ age of the more precise of two concordant and overlapping zircon fractions. Two titanite fractions also analyzed from the same sample are perfectly concordant and overlapping at a $^{206}Pb/^{238}U$ age of 183.0 ± 0.5 Ma. Although the zircon and titanite ages are within error and the somewhat younger age of titanite may be explained by its lower closure temperature, we take a more conservative approach in assigning a crystallization age of 183.2 ± 0.7 Ma based on the weighted mean $^{206}Pb/^{238}U$ age of the three most precise and concordant fractions.

The youngest stratigraphic unit intruded by the pluton is the Mount Brock volcanics (Evenchick and McNicoll 1993), the youngest volcanic member of the Hazelton Group in the Spatsizi area (Thorkelson et al. 1995). Critical fossil localities providing constraints on the age of the Mount Brock volcanics are found in intercalated marine sedimentary rocks (Read and Psutka 1990). Early and middle Toarcian ammonites are reported in all previous studies (Read and Psutka 1990; Evenchick and McNicoll 1993; Thorkelson et al. 1995). Our new fossil collections from the area west of Mount Brock indicate that volcanism started in latest Pliensbachian time (see also Thomson et al. 1986), and the evidence for middle Toarcian needs revision. *Dactylioceras*, commonly occuring in (but not restricted to) the lower Toarcian was found stratigraphically above the only collection that had suggested a middle Toarcian age in previous work, based on the identification of *Polyplectus* sp. As the ranges of *Polyplectus* and *Dactylioceras* are mutually exclusive, it is possible that the specimen in question actually represents some other, morphologically similar harpoceratid ammonite. In conclusion, we favour an interpretation that Mount Brock volcanism (and emplacement of the perhaps comagmatic McEwan Creek pluton) is not younger than early Toarcian.

Item 29 — Mount Brock Range

A late Early Jurassic volcanic unit within the Hazelton Group is informally known as the Mount Brock volcanics (Thorkelson et al. 1995). Northwest of Mount Brock, the volcanics interfinger with fossiliferous sedimentary rocks (Thomson et al. 1986). A felsic crystal tuff was U–Pb dated yielding an age of $180.4^{+8.0}_{-0.4}$ Ma (Pálfy et al. 2000*b*). The strongly asymmetric error results from a combination of $^{206}Pb/^{238}U$ ages of two concordant fractions and the mean $^{207}Pb/^{206}Pb$ age of all analyzed fractions. The sampled interval lies above a richly fossiliferous limestone bed of earliest Toarcian age. Earlier collections (Read and Psutka 1990) yielded *Dactylioceras* cf. *commune*, *D.* cf. *pseudocommune*, *D.* cf. *simplex*, and *D.* cf. *kanense* of the Kanense Zone (Jakobs 1992; Jakobs et al. 1994). Fossiliferous sedimentary intercalations become sparse within a thick sequence of andesitic volcanic rocks that overlie the dated tuff. Two fossil collections are used to provide an upper age bracket. One contains bivalves and an ammonite originally identified as *Polyplectus* sp., on which a Middle Toarcian age assignment was based (H.W. Tipper, personal communication, 1996). The other one, a new collection farther upsection, yielded *Dactylioceras* sp., whose range is mutually exclusive with *Polyplectus.* Assuming no structural repetition, this finding suggests an Early to early Middle Toarcian (Kanense to Planulata Zone) age for the entire succession, which can be reconciled with the first collection if its diagnostic ammonite is reinterpreted as an indeterminate harpoceratid.

Item 30 — Yakoun River

A volcanic ash layer near the base of the Middle Toarcian Crassicosta Zone exposed along the Yakoun River on Queen Charlotte Islands, British Columbia (Jakobs et al. 1994) was dated at 181.4 ± 1.2 Ma (Pálfy et al. 1997). The sample was obtained from the type section of this regional North American standard ammonoid zone, thus this date is considered a first rate calibration point. This U–Pb zircon date is based on analyses of single grains, as well as multi-grain fractions. Correlation with the Variabilis Zone of the primary standard northwest European zonal scheme is shown to be accurate with a margin of error not exceeding a subzone (Pálfy et al. 1997).

Item 31 — Diagonal Mountain 1

Diagonal Mountain in northwestern British Columbia is underlain by a thick succession of colour-banded fine clastics of the Hazelton Group (Evenchick and Porter 1993). Ammonite biostratigraphy of the area is subject of an ongoing study by G. Jakobs (Jakobs 1993). Several tuff layers were sampled for U–Pb zircon geochronology from measured sections with good ammonoid biostratigraphic control (Pálfy et al. 2000*b*). An altered ash layer from above the occurrence of *Yakounia* cf. *silvae* and *Pleydellia* (suggesting the uppermost Toarcian Yakounensis Zone) was dated as 179.8 ± 6.3 Ma. The interpretation is based on the weighted mean ^{207}Pb/^{206}Pb age calculated from three slightly discordant fractions exhibiting Pb loss. Early Bajocian sonninid ammonites were recovered above the sampled tuff.

Item 32 — Julian Lake dacite

A thick pile of felsic volcanics with locally interbedded marine sedimentary rocks occur in the Salmon River Formation near Julian Lake (Iskut River area, northwestern British Columbia). A dacite flow near its base yielded a U–Pb zircon age of 178 ± 1 Ma (J. K. Mortensen, unpublished data, 1996 and P.D. Lewis, personal communication, 1996). Hyaloclastite observed at the flow top points to submarine emplacement, in turn suggesting that the sedimentary rocks are not significantly younger than the dated flow that they directly overlie. These volcanic sandstones are locally fossiliferous and yielded a diverse ammonoid fauna (*Yakounia silvae*, *Pleydellia* cf. *maudensis*, *P.* cf. *crassiornata*, *Phymatoceras* sp.) clearly indicating the uppermost Toarcian *Yakounensis* Zone (J. Pálfy, unpublished data, 1996).

Several hundreds of metres upsection, another dacite flow yielded an age of 172.3 ± 1.0 Ma (J.K. Mortensen, unpublished data, 1996 and P.D. Lewis, personal communication, 1996). No identifiable fossils have been found in the upper part of the section, therefore the latter date can only be used with a latest Toarcian lower bracket and a Bajocian upper bracket, inferred from regional geology as the age of cessation of Salmon River felsic volcanism.

Item 33 — Treaty Ridge

The Treaty Ridge section is an important reference section for the Hazelton and overlying Bowser Lake groups in the Iskut River map area (northwestern British Columbia) (Lewis et al. 1993). Four separate samples from the upper felsic unit of the Salmon River Formation have been recently dated by the U–Pb method. Complex isotopic systematics that suggest presence of Late Triassic xenocrystic zircon and (or) Pb-loss hampered interpretation of data for one sample (V. McNicoll and R.G. Anderson, personal communication, 1995). Only one sample yielded concordant analyses giving an interpreted age of 177.3 ± 0.8 Ma (R.M. Friedman and R.G. Anderson, personal communication, 1996). A weighted mean ^{207}Pb/^{206}Pb age of two discordant but apparently inheritance-free fractions gives a 178 ± 12 Ma age for another sample (R.M. Friedman and R.G. Anderson, personal communication, 1996).

Fossiliferous sediments yielded ammonites both below and above the felsic volcanic unit (Lewis et al. 1993; Nadaraju 1993; Jakobs and Pálfy 1994). The Upper Aalenian *Howelli* Zone is documented by the presence of *Erycitoides* cf. *howelli*, *Pseudolioceras* cf. *whiteavesi*, *Tmetoceras* cf. *kirki*, and *Leioceras*? sp. in the underlying mudstone. *Sonninia*? sp., *Stephanoceras* sp., and *Zemistephanus* sp., collected from siltstone overlying the dated volcanic unit, assign an Early Bajocian upper age limit to the volcanism.

Items 34–35 — Eskay rhyolite

A flow-banded rhyolite unit that underlies the locally mineralized argillite at the Eskay Creek gold mine (Iskut River area, northwestern British Columbia) was sampled for U–Pb zircon dating on the east limb of the Eskay anticline and yielded an age of 174^{+2}_{-1} Ma (Childe 1996). *Erycitoides* cf. *howelli*, the index ammonite of the Late Aalenian Howelli Zone, was collected from above the rhyolite on the same limb of the anticline, less than 3 km away from the U–Pb sampling site. The rhyolite is therefore Upper Aalenian or older. No precise lower bracket can be assigned. The latest Pliensbachian (or possibly earliest Toarcian) ammonites listed for the Eskay porphyry (item 27) are the youngest fauna from underlying units in the area. Regionally, felsic volcanism correlative to the Eskay rhyolite is not known to begin prior to the latest Toarcian (see item 32).

Another sample from the same rhyolite unit on the west limb of the Eskay anticline yielded an age of 175 ± 2 Ma (Childe 1996). No ammonites have been found near this sample site but radiolarians identified from drill core obtained from the overlying argillite are dated as Aalenian to possibly early Bajocian (Nadaraju 1993). The local correlation of the rhyolite and argillite units is well documented by detailed geological mapping and further supported by the statistically indistinguishable U–Pb dates (Childe 1996) and concordant ammonoid and radiolarian ages. Therefore the same stratigraphic constraints are applied to both isotopic dates.

Item 36 — Diagonal Mountain 2

Another sample from Diagonal Mountain (see item 31 for description and references to local geology), a clay-rich tuff layer yielded an imprecise U–Pb date hampered by scarcity of zircon and Pb loss. The best age estimate of $167.2^{+10.5}_{-0.4}$ Ma results from a combination of ^{206}Pb/^{238}U age of the concordant fraction with the more conservative three-point weighted mean ^{207}Pb/^{206}Pb age of all fractions (Pálfy et al. 2000*b*). A fossiliferous bed below the zircon sample contains *Leptosphinctes* (*Prorsisphinctes*) cf. *meseres*, *L.* cf. *cliffensis*, and *Stephanoceras* sp. (G. Jakobs, personal communication, 1996). Elsewhere in the section, in which ammonites occur only sparsely, poorly preserved sonninids and stephanoceratids were found. The age of the sampled tuff layer is early Late Bajocian based on the co-occurrence of perisphinctids and stephanoceratids, characteristic of the Rotundum Zone (Hall and Westermann 1980).

Item 37 — Gunlock

The Middle Jurassic Carmel Formation in Utah contains numerous ash beds several of which have been recently dated using the ^{40}Ar/^{39}Ar single crystal laser probe method (Kowallis et al. 1996). One of them is published in detail and thus included in our database: a 166.3 ± 0.8 Ma sanidine age from the upper part of the Carmel Formation near Gunlock (Kowallis et al. 1993). In the chronogram calculation we use a ±3.5 Ma external error, which takes into account the decay constant uncertainties and systematic biases inherent in the ^{40}Ar/^{39}Ar ages. Characteristic bivalves allow correlation of the Carmel Formation. with the more fossiliferous Sliderock and Rich members of the Twin Creek

Limestone that yielded ammonites of late Early to Late Bajocian age (Imlay 1967, 1980).

Younger K–Ar (161.2 ± 1.8 Ma (1σ)) and Rb–Sr (162 ± 6.5 Ma (1σ)) ages from the Carmel Formation were reported as NDS102a and NDS102b and also used in GTS89.

Item 38 — Burnaby Island Plutonic Suite

Plutonic rocks assigned to the Burnaby Island Plutonic Suite (BIPS) yielded several U–Pb dates from the Queen Charlotte Islands. They range in age between 172 and ≥158 Ma (Anderson and Reichenbach 1991). We use here the four dates (Poole Point: 168^{+4}_{-1} Ma; Shields Bay: 168 ± 4 Ma; Rennell Sound: 164^{+4}_{-2} Ma (Anderson and Reichenbach 1991); and Cumshewa Head: 167 ± 2 Ma (Anderson and McNicoll 1995)), which form a tight cluster in the older part of the age range of the BIPS. The youngest country rocks crosscut by BIPS intrusions are Bajocian volcano-sedimentary strata of the Yakoun Group. Ammonoid biochronology of the Yakoun Group elsewhere in the Queen Charlotte Islands suggests an age range of Widebayense to Oblatum zones (Hall and Westermann 1980; Poulton et al. 1991b). It is possible that the Yakoun Group volcanics represent extrusive equivalents of BIPS plutons. In the future, isotopic dating of the Yakoun Group volcanics holds promise for contributing to better constraints on the Bajocian time scale. At present, we use a pooled age of 168 ± 4 Ma for early BIPS plutons as a minimum constraint for the undivided Bajocian stage. The late Bathonian and younger age of the Moresby Group (Poulton et al. 1991b) which disconformably overlies the Yakoun Group and is not known to be crosscut by BIPS is used to provide a minimum age for the BIPS, although the oldest strata found directly deposited on BIPS rocks are Early Cretaceous in age (Anderson and Reichenbach 1991).

Item 39 — Harrison Lake

A U–Pb zircon age of 166.0 ± 0.4 Ma was reported from a rhyolite near the top of the Weaver Lake Member (Harrison Lake Formation) exposed on Echo Island in Harrison Lake, southwestern British Columbia (Mahoney et al. 1995). No age diagnostic fossils have been recovered from sedimentary interbeds within the Weaver Lake Member (Arthur et al. 1993). The youngest fossil known from the underlying Francis Lake Member is *Erycitoides*? sp., and *Tmetoceras scissum* of Late Aalenian age. It has been suggested that the Weaver Lake Member is as young as Early Bajocian or possibly even younger (Arthur et al. 1993; Mahoney et al. 1995). The Harrison Lake Formation is unconformably overlain by the Mysterious Creek Formation which yielded a rich Early Callovian ammonite fauna (Arthur et al. 1993). The unconformity separating the units represents a regional deformation event and strongly suggests that the Weaver Lake Member is significantly older than Early Callovian.

The isotopic age is further supported by another two, nearly identical U–Pb ages from related rocks in the area. A rhyolite dyke from strata correlative to the upper part of the Weaver Lake Member near the Seneca mineral deposit yielded an age of $165.9^{+6.4}_{-0.3}$ Ma (McKinley 1996) and a comagmatic quartz-feldspar porphyry stock (Hemlock Valley stock) was dated at 166.0 ± 0.4 Ma (Mahoney et al. 1995).

Item 40 — McDonell Lake

Near McDonell Lake, northwestern British Columbia, Bajocian to Oxfordian fossiliferous strata are known to occur (Frebold and Tipper 1973). Within a section of volcanogenic sandstone, a strongly reworked tuff or tuffaceous sandstone was sampled. Based on two samples taken from adjacent beds that have no demonstrable difference in biochronologic age, a best age estimate of $158.2^{+1.9}_{-0.4}$ Ma was obtained (Pálfy et al. 2000*b*). *Iniskinites* sp., and perisphinctids occur both immediately below and above the isotopically dated beds. Ammonites recovered from above the tuff include *Kepplerites* ex gr. *tychonis*, *Kepplerites* sp, *Lilloettia* cf. *lilloetensis*, and *Xenocephalites* cf. *vicarius*. The assemblage best agrees with Fauna B6 of Callomon (Callomon 1984). *K.* ex gr. *tychonis* allows correlation with the basal Upper Bathonian where the genus first appears.

Item 41 — Copper River

Near Copper River (northwestern British Columbia), richly fossiliferous shallow marine sandstone of the Ashman Formation was deposited during the Bathonian–Callovian (Tipper and Richards 1976). A reworked volcanic ash sample yielded a small amount of zircon, which was dated as $162.6^{+2.9}_{-7.0}$ Ma (Pálfy et al. 2000*b*). The imprecise age estimate is based on a marginally concordant fraction (probaility of concordance = 0.15, calculated using the method of Ludwig, 1988) and a lower intercept age calculated using another fraction with Proterozoic inheritance. Such two-point discordia lines are not considered to provide a robust age estimate as mild Pb-loss may have effected both analyzed fractions. The ammonoid assemblage, including *Cadoceras* cf. *moffiti*, *Iniskinites* cf. *martini*, *Xenocephalites* cf. *vicarius*, “*Choffatia*” sp., *Kepplerites* ex gr. *loganianus*, and *Lilloettia* cf. *lilloetensis* indicates a Late Bathonian – Early Callovian age.

Item 42 — Diagonal Mountain 3

Another sample from Diagonal Mountain in northwestern British Columbia (see item 31 for description and references to local geology), from a tuff layer within the Ashman Formation, yielded a lower intercept age of 158.4 ± 0.8 Ma (Pálfy et al. 2000*b*). This date is interpreted as a minimum estimate of the unit’s age, because the effect of Pb loss could not be fully assessed due to the small amount of zircon available. Stratigraphic relationships of the sampled unit are obscured by tight folding. *Iniskinites* sp. occurs both below and above the zircon sample. Earlier collections yielded *Iniskinites* cf. *robustus* below and *Adabofoloceras* or *Lilloettia* above the tuff (G. Jakobs, personal communication, 1996). *Iniskinites* is taken to indicate the Late Bathonian as the most likely age of the sample although precise correlation around the Bathonian–Callovian transition is difficult (Callomon 1984; Poulton et al. 1994).

Items 43–44 — Chacay Melehué

Two tuff layers from the Chacay Melehué section in the Neuquén Basin, Argentina, were dated at 160.5 ± 0.3 Ma and 161.0 ± 0.5 Ma by zircon U–Pb method (Odin et al. 1992). Both dates are lower intercept ages, the first one is corroborated by one concordant fraction. The section has a well documented ammonite fauna that is dominated by endemic South American forms but allows correlation with the European standard. The lower sample is very near to the boundary of the regional Steinmanni and Vergarensis zones that is equated to the Bathonian–Callovian boundary (Riccardi et al. 1991). The upper sample was collected near the boundary between the Bodenbenderi and Proximum zones that approximately corresponds to the middle of the Boreal Calloviense Zone or the Mediterranean Gracilis Zone in the upper part of the Lower Callovian (Riccardi et al. 1991).

Item 45 — Tsatia Mountain

Lenses of boulder conglomerate within the Bowser Lake Group in the Bowser Basin (northwestern British Columbia) locally contain dacite clasts, one of which was dated by U–Pb zircon method. A precise age of 160.7 ± 0.7 Ma is based on analysis of three concordant fractions (Ricketts and Parrish 1992). The dated clast was recovered from a biostratigraphically well constrained section on Tsatia Mountain. An Early Callovian *Cadoceras* fauna was found below the conglomerate whereas overlying deposits yielded *Stenocadoceras* of Middle Calovian age and still younger assemblages higher upsection (Poulton et al. 1991a; Poulton et al. 1994). The dacite clast therefore cannot be younger than Middle Callovian.

Item 46 — Josephine ophiolite

U–Pb ages from plagiogranites in the Josephine ophiolite have been used in previous time scales. GTS89 (item HS1) uses an age of 157 ± 2 Ma as Oxfordian or older. The original data (Saleeby et al. 1982) consist of analyses of two different samples (one single-fraction and one based on two fractions of unabraded zircons). A later revision of one of these dates to 162 ± 1 Ma (Saleeby 1987) is not considered in GTS89 or MTS94 (item 253, labelled as HS2). A full documentation is now available (Harper et al. 1994), showing that the interpreted age is the $^{206}Pb/^{238}U$ age (with 1σ error) of the subsequently analyzed, concordant, abraded fraction. We assign a more conservative age estimate to samples when no duplicate concordant fraction is available and there is evidence of Pb-loss. A weighted mean $^{207}Pb/^{206}Pb$ age from the two fractions (sample A88Z) combined with the $^{206}Pb/^{238}U$ age of the concordant one as a minimum age gives 162^{+7}_{-2} Ma as the crystallization age. Furthermore, a U–Pb age of 164 ± 1 Ma was reported from the Devils Elbow ophiolite remnant thought to be correlative to the Josephine ophiolite (Wright and Wyld 1986).

The Josephine ophiolite is overlain by the Galice Formation, which yielded bivalves (*Buchia concentrica*), perisphinctid ammonites, and radiolarians (e.g., *Mirifusus*), indicating an Oxfordian age (Imlay 1980; Pessagno and Blome 1990).

Item 47 — Rogue Formation tuff breccia

A U–Pb zircon age of 157 ± 2 Ma (1σ) was obtained from tuff breccia in the Rogue Formation (Klamath Mountains, northern California) (Saleeby 1984; Harper et al. 1994). Analytical data have not been published. Considering the scarcity of Late Jurassic data, we include this date with a qualifier as a minimum age. It is warranted as Pb-loss is documented as a common phenomenon affecting the U–Pb systematics of Jurassic zircons from the area (Saleeby 1987). The Rogue Formation is overlain by the sparsely fossiliferous Galice Formation which yielded bivalves (*Buchia concentrica*) and radiolarians (e.g., *Mirifusus*) indicating an Oxfordian age (Imlay 1980; Pessagno and Blome 1990).

Item 48 — Hotnarko volcanics

A preliminary U–Pb age of 154.4 ± 1.2 Ma was obtained from a sample from the Hotnarko volcanics in the eastern Coast Belt in west-central British Columbia (van der Heyden 1991). Volcaniclastic rocks near the base of the succession yielded *Anditrigonia* aff. *plumasensis* (identification by T. Poulton), which ranges from the Callovian through the Middle Oxfordian. It thus provides a lower stratigraphic bracket for the dated volcanic rocks whereas geologic evidence suggested that the entire succession may have been deposited within a brief volcanic episode (van der Heyden 1991).

Item 49 — Tidwell Member (Morrison Formation)

Three sanidine $^{40}Ar/^{39}Ar$ single-crystal laser fusion ages (154.8 ± 1.2, 154.8 ± 1.1, and 154.8 ± 2.8 Ma) and one plateau age (154.9 ± 1.0 Ma) are reported from an ash bed in the Tidwell Member (lower part of the Morrison Formation) sampled in two sections of Utah (Kowallis et al. 1998). All analyses determine the age of the same ash bed and are remarkably concordant therefore we use the most precise of them in our database. In the chronogram calculation, we use a ±3.5 Ma external error, which takes into account the decay constant uncertainties and systematic biases inherent in the $^{40}Ar/^{39}Ar$ ages. The youngest cardioceratid ammonites from the underlying Redwater Shale (Stump Formation) are middle Oxfordian (Imlay 1980; Callomon 1984). The base of the Morrison Formation marks a regional unconformity. Ostracods and charophytes (Schudack et al. 1998), as well as palynomorphs (Litwin et al. 1998), suggest that the lower part of the Morrison Formation, including the Tidwell Member, is Kimmeridgian (likely Early Kimmeridgian) in age. Magnetostratigraphic studies also reveal correlation between the reversal sequence in the Morrison Formation and the Kimmeridgian oceanic magnetic anomaly pattern (Steiner et al. 1994).

Items 50–55 — Brushy Basin Member (Morrison Formation)

Six $^{40}Ar/^{39}Ar$ laser fusion ages for plagioclase from volcanic ash layers in the Brushy Basin Member (Colorado Plateau) range between 153 and 148 Ma (Kowallis et al. 1991). This dataset is now superseded by seven more precise $^{40}Ar/^{39}Ar$ sanidine laser fusion ages from different ash layers in five section of the upper part of the Brushy Basin Member (Kowallis et al. 1998). Six of them (150.3 ± 0.5, 150.2 ± 1.0, 149.3 ± 1.1, 149.3 ± 1.0, 149.0 ± 0.8, and 148.1 ± 1.0 Ma) are consistent with their relative stratigraphic position within the unit and are preferred. In the chronogram calculation, we use their external errors (ranging between ± 3.0–3.6 Ma), which take into account the decay constant uncertainties and systematic biases inherent in the $^{40}Ar/^{39}Ar$ ages. All of the dated ash layers occur in the upper part of the unit, above the level marking a characteristic change in clay mineralogy (Kowallis et al. 1998).

Correlation based on ostracods and charophytes suggests the isotopically dated interval to be Kimmeridgian in age in its lower part, possibly ranging to the Tithonian near the top (Schudack et al. 1998). Palynological results are less conclusive but corroborate the placement of these strata to the Kimmeridgian and Tithonian (Litwin et al. 1998). We infer an age restricted to the Kimmeridgian for the lowest two dated ashes.

Item 56 — Grindstone Creek tuff

A precise and well documented U–Pb zircon age of $137.1^{+1.6}_{-0.6}$ Ma was obtained from two altered crystal tuff layers within a fossiliferous Upper Jurassic to Lower Cretaceous mudstone succession (Great Valley Sequence) in California by (Bralower et al. 1990). The dated levels are assigned to *Buchia uncitoides* and *B. pacifica* bivalve zones and the *Assipetra infracretacea* (NK–2A) nannofossil subzone, both indicating a Late Berriasian age. Correlation with Deep Sea Drilling Project (DSDP) sites containing integrated magnetostratigraphic and nannofossil records permited their assignmnent to the CM16 or CM16n magnetochrons (Bralower et al. 1990). A revision of the ranges of critical nannofossil taxa suggests that the dated level can represent the CM16–CM15 magnetochrons, still within the Upper Berriasian (Channell et al. 1995). This date is used in MTS94 as item 232.

Appendix 2.

Comments on items used in earlier time scales, but rejected in this study

Item HLB1 (GTS89) and item 269 (MTS94)

The quoted $^{40}Ar/^{39}Ar$ age is 185.0 ± 3.0 Ma (1σ) with Pliensbachian–Toarcian stage brackets. The original source (Hess et al. 1987) reports 11 ages, ranging from 190 to 180 Ma, from a 2000 km^2 area in the northern Caucasus underlain predominantly by thick volcanosedimentary sequences. No detailed stratigraphy was reported but sedimentary rocks generally associated with the volcanic rocks were said to contain rare Pliensbachian bivalves and brachiopods, as well as Toarcian ammonites. We conclude that the coarse stratigraphic resolution of this data set does not warrant its inclusion in the time scale calibration and the simple averaging of the 11 different ages is inadequate for inclusion in our database.

Item 251 (MTS94)

MTS94 quotes an age of 155.3 ± 3.4 Ma as an $^{40}Ar/^{39}Ar$ age for Oxfordian oceanic crust recovered from DSDP site 765. However, it is clear from the original source (Ludden 1992) that the age in question was obtained using the K–Ar method, therefore we exclude it from our database.

Item 258 (MTS94)

Gradstein et al. (1994) quote an $^{40}Ar/^{39}Ar$ age of 166.8 ± 4.5 Ma obtained by Pringle (1992) as an age for Callovian oceanic crust recovered from DSDP site 801. According to Pringle (1992), this is the age of tholeiitic mid-ocean ridge basalts (MORB) basalt recovered from the base of the drillhole. It is overlain by alkaline off-ridge basalt that yielded an $^{40}Ar/^{39}Ar$ age of 157.4 ± 0.5 Ma. Magnetostratigraphic evidence suggests that the site was drilled into the Jurassic Quiet Zone some 450 km away from anomaly M37, the oldest known preserved oceanic magnetic lineation thought to be Callovian in age. At DSDP site 801, original radiolarian biostratigraphy from the oldest overlying sedimentary rocks indicated a latest Bathonian to earliest Callovian age (Matsuoka 1992). This interpretation was challenged by Pessagno and Meyerhoff Hull (1996) who considered the fauna as Middle Oxfordian and argued that the younger isotopic age supports this assignment. Considering the controversy that surrounds this dataset, we chose not to include it in the present compilation.

NDS184 (revised in GTS89) and item 272 (MTS94)

Both GTS89 and MTS94 quoted an $^{40}Ar/^{39}Ar$ age of 197.6 ± 2.5 Ma from the Toodoggone volcanics in British Columbia, based on an undocumented revision of NDS184 (Armstrong 1982). The stratigraphic age given is Sinemurian to Toarcian but the date is rejected on the basis of a lack of analytical documentation. Clark and Williams-Jones (1991) recently obtained a fully documented $^{40}Ar/^{39}Ar$ age of 195.1 ± 1.6 from the Toodoggone volcanics but there is no fossil data available to bracket the age of these mainly subaerial volcanic rocks. Although it can be reasonably correlated with the early phase of Hazelton Group volcanism recorded elsewhere (e.g., Telkwa Formation, Cold Fish volcanics), it is unuseable for time scale calibration.

Item SCH1 (GTS89) and item 245 (and 247?) (MTS94)

GTS89 used a U–Pb age of 153 ± 1 Ma in the Kimmeridgian. MTS94 revised the error to ± 3 Ma and lists apparently the same age twice, the second time with Oxfordian–Kimmeridgian brackets. In fact this is not a U–Pb age itself but rather a “conservative estimate of the age of the Nevadan deformation” (Schweikert et al. 1984). It is based on several, poorly documented or single-fraction early determined U–Pb ages pooled together with K–Ar dates. These items are rejected as a composite of questionable veracity and validity to time scale calibration.

Item HS2 (GTS889) and item 244 (labelled as HS1 in MTS94)

GTS89 quoted a U–Pb age of 150.5 ± 2 Ma as Kimmeridgian or younger. Evidently the quoted date derived from a combination of two single-fraction U–Pb ages of unabraded zircons (Saleeby et al. 1982; Harper 1984): a 151 ± 3 Ma age obtained from a dyke intruding the Josephine ophiolite and a second 150 ± 2 Ma age from a sill intruding the overlying Galice Formation. These ages are rejected as suspect because Pb-loss is documented from several other samples from the same dataset in subsequent work on abraded zircons (Saleeby 1987; Harper et al. 1994).

Item HMP2 (GTS9) and item 239 (MTS94)

The Tithonian or older U–Pb age of 152.5 ± 2 Ma, used in GTS89 and MTS94, presumably represents a mean of $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ ages reported by Hopson et al.

(1981). This date doesn't satisfy our criteria for inclusion in the database, as it is based on the analysis of a single fraction.

Item HMP3 (GTS9) and item 240 (MTS94)

The Tithonian or older U–Pb age of 154 ± 2 Ma used in GTS89 and MTS94 is apparently the $^{207}Pb/^{235}U$ age of a slightly reversely discordant fraction reported by Hopson et al. (1981). We reject this date as being based on the analysis of a single fraction, where reverse discordance likely indicates analytical problems.

Item HMP4 (GTS89) and item 255 (MTS94)

The quoted Oxfordian or older U–Pb age is 162 ± 2 Ma, apparently the younger $^{206}Pb/^{238}U$ age of the two unabraded fractions analyzed by Hopson et al. (1981). This fraction is slightly reversely discordant and the other fraction yielded a $^{207}Pb/^{206}Pb$ age of 184 ± 10 Ma, which is outside the error of the $^{206}Pb/^{238}U$ age of 165 ± 2 Ma. This discrepancy of apparent ages points to complexities in the U–Pb systematics. This date is rejected as the true crystallization age cannot be unambiguously resolved from the published analyses.

Item HMP1 (GTS89) and item 254 (MTS94)

The quoted Oxfordian or older U–Pb age is 161 ± 2 Ma based on data of Hopson et al. (1981). Of the two analyzed unabraded fractions, one yielded a significantly older apparent $^{207}Pb/^{206}Pb$ age outside the error of the $^{206}Pb/^{238}U$ age. We interpret this as an indication of Pb-loss, inheritance, or both. On the same grounds as in the previous item, we reject this date, as the true crystallization age cannot be unambiguously resolved from the published analyses.

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