The global distribution, setting, and dynamic implications of Ordovician orogenesis are reviewed. Evidence for true Ordovician continent-continent collision is absent. Orogenesis is principally due to accretion of arc terranes and/or ribbon microcontinents. Most arc terranes are ensialic and separated from the adjacent continents by backarc basins, the episodic closure of which commonly was responsible for orogenesis. Little evidence is preserved for true intra-oceanic juvenile arcs during the Early to Middle Ordovician. Instead, subduction appears to have been localized near the margins of Laurentia, Gondwana, Baltica, and Siberia, forming extensive linear orogenic belts during relatively short periods when the upper plate switched from extension to compression. Such tectonic switching appears to have taken place along the entire length of the Pacific and Iapetan margins of Gondwana (>10,000 km) from Middle–Late Cambrian to Early Ordovician time. The onset of orogenesis along Gondwana’s Pacific margin during the end of the Early Cambrian (ca. 513 Ma) coincided with subduction initiation along both margins of the Iapetus Ocean. Orogenesis and subduction initiation are causally related to a global-scale plate reorganization, probably induced by terminal amalgamation of Gondwana. During the Paleozoic, Laurentia’s Iapetan margin steadily grew in size and expanded southward owing to continuous accretion of suprasubduction zone oceanic crust, peri-Gondwanan arc terranes, and ribbon microcontinents. In contrast, the Pacific, Iapetan, and Rheic margins of Gondwana saw little addition of new, allochthonous crust. Accretion mainly involves
reattachment of previously rifted-off arc terranes and small slivers of the intervening marginal
basin crust.

INTRODUCTION

Orogenic events (orogenesis) affected many parts of the world during the Ordovician, and the remnants of the created mountain belts created principally can be found principally in the present-day continents of North America, South America, Australia, Antarctica, and Eurasia. There is also minor Ordovician deformation in Africa, but this mainly relates to transpression and basin inversion induced by far-field stresses and is not relevant to this paper. What makes Ordovician orogenesis stand out from all other periods in the Paleozoic is that nowhere it appears to have involved true continent-continent collision. Some workers have invoked a Middle Ordovician collision between Gondwana and Laurentia (e.g., Dalziel et al., 1994), mainly based on orogenesis taking place nearly coevally on both margins of Iapetus (Fig. 1) and involving accretion of Laurentian-derived crust (Precordillera) to Gondwana. However, this model was abandoned by Dalziel (1997) and replaced by a short-lived collision between a large promontory on Laurentia that bulged into the Iapetus Ocean (Texas Plateau) and Gondwana. Most other workers favor collision between a Laurentian-derived microcontinent (Cuyania/Precordillera) and Gondwana instead (e.g., Ramos et al., 1986; Astini et al., 1995; Thomas and Astini, 1996). Irrespective of whether or not the fleeting promontory-Gondwana collision ever happened, we do not consider it a true continent-continent collision, because the area involved in this putative collision was very small compared with the length of the margins. Instead the four principal continental masses present on Earth at that time (Laurentia, Gondwana, Baltica, and Siberia; see Fig. 1 and also Cocks and Torsvik, 2002; Torsvik and Cocks, 2005) were involved in accretion of peri-cratonic arc terranes and/or ribbon-shaped microcontinents and remained separated by oceans (principally the Iapetus, Aegir, Tornquist-Ran, and Pacific Oceans). These accretionary events were precursors to major continent-continent collisions that occurred during the middle to late Paleozoic. Formation and accretion of ribbon-shaped microcontinents, principally derived from Gondwana and to a lesser extent Laurentia, and the closure of marginal oceanic basins on a global scale during the Ordovician are important tectonic phenomena that pose important geodynamic questions.

The purpose of this paper is to provide a brief review and summary of the data and ideas related to Ordovician orogenic events, mainly concentrating on their tectonic evolution and whether they are kinematically and/or dynamically linked, and if so, how and why, and finally to discuss the implications of these orogenic events. The latter is done to further our understanding of the global tectonic climate and the driving tectonic forces during the Ordovician.

In this paper we have adopted the usage of the new Ordovician stage names (Finney, 2005; Bergström et al., 2006).

OROGENESIS AND TIME

In general, classification and correlation of structures and other evidence of tectonic events were historically (pre–plate tectonics) done on the basis of time alone, which is a hazardous process (although one sometimes has no other choice), because orogenic events are invariably diachronous, and kinematically unrelated tectonic events may take place coevally in different parts of the world (e.g., van Staal, 1994). These, by chance, may have been juxtaposed later as a result of subsequent plate motion(s) and closure of intervening oceans (van Staal et al., 1998).

ORDOVICIAN OROGENESIS
Orogenic events discussed in this paper include the Finnmarkian, Penobscottian,
Taconian, Grampian, Shelveian, and the early stages of the Salinic and Scandian in the
Appalachians and Caledonides of North America, British Isles, and Scandinavia; M’Clintock in
Pearya of Arctic Canada and its unnamed equivalent in neighboring Svalbard; Famatinian and/or
Ocloyic in South America; Sardic or Eo-Variscan in the Eurasian Variscan orogen; events in the
Central Asian orogenic belt (Altaids); and parts of the Ross-Delamerian and Lachlan orogenies
in eastern Australia and Antarctica (Fig. 1). Most of these orogenies are restricted to the
Ordovician, but some started or mainly took place during the Middle to Late Cambrian
(Finnmarkian and Ross-Delamerian), whereas others started during the latest Ordovician and
climaxed during the Silurian (Salinic and Scandian). Their inclusion here is either due to
diachroneity, such that their orogenesis locally extends into the Ordovician, and/or to their
dynamics having had an impact on understanding Ordovician orogenesis in general. Following
the definition of McKerrow et al. (2000), we restrict usage of the term **Caledonian orogeny** to
describe all orogenic events involved in closing the Iapetus Ocean.

**Ordovician Paleogeography**

Comprehensive reviews of Earth’s geography, position of tectonic elements, and
migration of terranes relevant to this paper are presented in van Staal et al. (1998), Cocks and
Although these authors do not necessarily agree on the evolution of each element, they broadly
agree on the disposition of the main continental landmasses summarized in Figure 1. We adhere
to these views, which are significantly different from the one proposed by Dalziel (1997).

**North American Appalachians and British Caledonides**

Ordovician orogenesis affected the Appalachians differently in different parts of the
mountain chain. These orogenic events generally are diachronous along the length of the
mountain chain, and to make matters more confusing, they occurred on both sides of the Iapetus
ocean while the ocean was still very wide (Fig. 1), which makes the classification of Ordovician
structures difficult without knowing the provenance and tectonic setting of the deformed rocks
involved. The oldest orogenic events that affected the Appalachian orogen are represented by the
latest Cambrian–Early Ordovician phases of the Taconian on the Laurentian side and the
Penobscottian on the Gondwanan side (e.g., Colman Sadd et al., 1992; van Staal et al., 2007).
Early phases of both orogenies partly overlap in time, and both involve ophiolite accretion-
obduction and arc-continent collision (Fig. 1). The onset of ocean closure along opposite margins
of the Iapetus Ocean (Fig. 1) signals a major plate reorganization at the end of the Early
Cambrian (van Staal et al., 1998; also see below).

The main phase of the Taconian (Figs. 1 and 2) in the northern Appalachians is the
collision between magmatic arc(s) and Laurentia, which started during the Early Ordovician
(Humberian in Newfoundland), and lasted until the Late Ordovician. Deformation and
accompanying tectonothermal activity are principally related to collision of a magmatic arc
(Notre Dame–Shelburne Falls arc) built upon a ribbon-shaped peri-Laurentian **microcontinent**
(Karabinos et al., 1998) named **Dashwoods** in Newfoundland (Taconic 2 suture in Fig. 1;
Waldron and van Staal, 2001; van Staal et al., 2007), rapidly followed by Late Ordovician (ca.
450 Ma) soft-docking of the peri-Gondwanan Popelogan–Victoria–Bronson Hill arc (Taconic 3,
Fig. 2) built on a sliver of Ganderian crust (van Staal, 1994; Rogers et al., 2006). This phase
represents the terminal phase of the Taconian and closed the main oceanic tract of the Iapetus
between Laurentia and Ganderia (Fig. 2), although a wide oceanic **backarc** basin (Tetagouche-
Exploits backarc) and seaway (Acadian seaway) remained open, to be closed during the Silurian.
and Early Devonian, respectively (van Staal et al., 1998, 2007; van Staal, 2005; Valverde-Vaquero et al., 2006).

In the southern Appalachians the Taconian is termed Blountian (Kay, 1942; Rodgers, 1971; Drake et al., 1989). Orogenesis here is particularly evident from the accumulation of the substantial Blount-Sevier foredeep clastic wedge on Laurentia’s continental margin near the beginning of the Middle Ordovician (ca. 468 Ma, Darriwilian), but it is also recorded by abundant metamorphic ages, formation of eclogite, and generation of arc-like plutons in the internal parts of the mountain chain (central and eastern Blue Ridge and western Inner Piedmont). The most attractive model, which is virtually identical to the northern Appalachians, is one involving formation of a west-facing magmatic arc on a peri-Laurentian microcontinent (Hatcher, 1989; Miller et al., 2006; Hibbard et al., in press). Metamorphism as high as granulite facies affected the rocks in the internal parts of the southern Appalachians between 455 and 465 Ma (Moecher et al., 2004) as a result of the arc-continent collision and associated obduction. Hibbard (2000) proposed that immediately subsequent to this collision the leading edge of the peri-Gondwanan Carolina terrane docked in the latest Ordovician–Early Silurian. If Hibbards model is correct, it indicates a remarkably similar kinematic/dynamic evolution for the entire Appalachian margin of Laurentia (Fig. 2) during the Ordovician (Hibbard et al., in press), virtually eliminating the possibility of a fleeting Middle–Late Ordovician Gondwana-Laurentia collision as proposed by Dalziel et al. (1994) and Dalziel (1997). Other compelling sedimentologic, geochronologic, and field data from the southern and central Appalachians also support closure of the bulk of the Iapetus Ocean in the Late Ordovician, but these data also indicate that the trailing Rheic Ocean remained open until its late Paleozoic destruction (Bream, 2003; Merschat et al., 2005; Hatcher and Merschat, in press). Extensive peri-cratonic island arc systems fringing both the Iapetan margins of Laurentia and Gondwana until the Late Ordovician (van Staal et al., 1998), combined with paleomagnetic and faunal evidence (e.g., Cocks and Torsvik, 2002), require that both continents remained widely separated until at least the Late Devonian. Hence, we conclude that there is no evidence to support collision of Laurentia with any large continental landmass other than ribbon-shaped continental arc terranes during the Ordovician.
The Taconian is kinematically and temporally equivalent to the Grampian in the British Isles (Dewey and Shackleton, 1984). Like the Taconian, the Grampian starts with ophiolite obduction immediately prior to arc-continent collision in the Early Ordovician (Dewey and Mange, 1999). The main Grampian orogenesis is a short-lived ca. 15 Ma event (ca. 475–460 Ma), although it comprises a complex tectonometamorphic cycle involving early blueschist and eclogite facies metamorphism of the subducting Laurentian (Dalradian) margin (Friend et al., 2000; Chew et al., 2003), polyphase deformation, and medium to high grade Barrovian metamorphism of the collision complex concurrent with syntectonic intrusion of plutons (Friedrich et al., 1999; Soper et al., 1999; Dewey and Mange, 1999). In this respect the Grampian is nearly coeval and kinematically remarkably similar to the main phase of the Taconian in central and western Newfoundland (van Staal et al., 1998, 2007). Grampian orogenesis was rapidly followed by the Late Ordovician–Early Silurian Scandian orogenesis similar in nature to the Salinic in the northern Appalachians. While Grampian orogenesis occurred on the Laurentian side, Early to Middle Ordovician orogenesis also took place on the Gondwanan side in the British Isles (e.g., Max et al., 1990; Todd et al., 2000). This orogenic event, which has no name attached to it as yet in the British Isles, is the equivalent of the Penobscottian in the Appalachians (van Staal et al., 1998).

Scandinavian, Greenland, and Svalbard Caledonides

Orodovician orogenesis in the Scandinavian Caledonides was, until recently, mainly thought to be represented by an early event that principally took place in the Late Cambrian along the Baltic margin (Finnmarkian; Andréasson et al., 1998; Sturt and Ramsay, 1999; Grenne et al., 1999; Roberts, 2003). However, the structurally highest allochthons, in the Scandinavian Caledonides, which have a Laurentian provenance, have preserved ample evidence for a major Ordovician tectonic event that took place prior to their Silurian (Scandian) docking with the Baltic margin (Yoshinobu et al., 2002; Roberts, 2003). The Ordovician event is kinematically equivalent to the Taconian–Grampian orogeny in the Appalachians and British Caledonides. Surprisingly, no evidence for this event has been preserved in the East Greenland Caledonides, the area from which these rocks presumably were derived. However, the presence of Middle Ordovician blueschist, eclogite, and other evidence of major tectonism in western Svalbard (Dallmeyer et al., 1990; Gee and Page, 1994) and Pearya (M’Clintock orogeny; Trettin, 1987) leave little doubt that the Taconian–Grampian was probably continuous along the entire length of Laurentia’s Iapetan margin north of the Appalachians. The lack of evidence of this event in East Greenland suggests that most Ordovician tectonism was restricted to the peri-Laurentian arc terrane preserved in the Caledonides upper allochthon (Hølanda terrane of Grenne et al., 1999), which is thought to have been present directly outboard of Laurentia at the latitude of Greenland (Fig. 1) during the Ordovician (Grenne et al., 1999). A similar relationship has been observed in Newfoundland, where most Taconian tectonothermal activity was restricted to the peri-Laurentian Notre Dame arc and its basement (the Dashwoods microcontinent of Waldron and van Staal, 2001) and largely absent in the adjacent passive margin (van Staal et al., 2007).

The Finnmarkian event is generally thought to be mainly restricted to the latest Cambrian, but age dating revealed that orogenesis continued into the Early–Middle Ordovician in Finnmark in northern Norway (e.g., Rice and Frank, 2003). The Finnmarkian involved arc development, ophiolite obduction, arc accretion, and metamorphism to eclogite facies conditions that principally affected rocks that now reside in the middle and upper allochthons in Norway and Sweden (Gee et al., 1985). The main Finnmarkian event is thought to have been the result of
collision of a peri-cratonic arc terrane with the adjacent Baltic margin (Sturt and Roberts, 1991; Grenne et al., 1999; Roberts, 2003). Holtedahl (1920) recognized a slightly younger Early Ordovician event in central Norway and called it the Trondheim event, which probably tectonically is related tectonically to Finmarkian orogenesis. It includes ophiolite emplacement and blueschist and eclogite facies metamorphism (Eide and Lardeaux, 2002; Roberts, 2003). Ophiolites have yielded uranium lead ages of 497–482 Ma (Dunning and Pedersen, 1988; Roberts, 2003), which are identical in age to the Penobscot ophiolites preserved in Newfoundland. In general, the close similarities in age and style of orogenesis between the Penobscottian and Finmarkian, both positioned on the southern side of Iapetus, invite correlation. If correct, Baltica and adjacent Gondwana (Ganderia and parts of Avalonia; Rogers et al., 2006) may have been characterized by a more or less continuous (excluding transform faults) peri-cratonic arc terrane with a length on the order of thousands of kilometers. Thus consistent to what has been observed in the British Caledonides and the northern Appalachians, Ordovician orogenesis also appears to have taken place on both sides of the Iapetus Ocean in the Scandinavian Caledonides.

The Baltic–East European craton (Baltica) also converged with Ganderia and Avalonia during closure of the Tornquist Ocean, which separated Baltica from Gondwana, along the former’s southern margin (Valverde-Vaquero et al., 2006), and which led to formation of the Thor suture. The Tornquist closure was completed by the end of the Ordovician (454–443 Ma) and caused the Shelveian orogeny, whose effects on the exposed parts of Baltica (lower plate) appear to be minor (Torsvik and Rehnstrom, 2003). In general, this collision appears to have been relatively soft, with deformation and metamorphism (low grade) mainly restricted to the collision zone (Trans-European suture zone [TESZ]; see below).

Scandian orogenesis in the Scandinavian Caledonides mainly took place during Silurian times. The Scandian was principally caused by the main collision between Laurentia (Greenland) and the Baltic craton. It is coeval and is correlated with the Salinic orogeny (Laurentia-Ganderia collision) in the northern Appalachians (van Staal, 2005). Since Baltica and Ganderia were already together by Late Ordovician times as a result of the Shelveian orogeny (see above and Valverde-Vaquero et al., 2006), the Salinic and Scandian orogenies are both dynamically related to collision of Laurentia (upper plate) and the composite craton comprising Baltica and Ganderia (lower plate).

## Early Paleozoic Orogenic Events in Eurasian Gondwana

The orogenic events that involved the assembly of Paleozoic Europe comprised kinematically and temporally diverse, and in part, unrelated events that were scattered throughout the Paleozoic, culminating in the collision that formed Pangea at the end of the Paleozoic. Orogenic events that took place during the Ordovician include the Shelveian orogeny, which formed as a product of convergence among Baltica, Ganderia, and Avalonia (McKerrow et al., 1991; Pharaoh et al., 1993; van Staal et al., 1998; Torsvik and Rehnstrom, 2003; Valverde-Vaquero et al., 2006). The structures related to the Shelveian orogeny are mainly present in the unexposed basement beneath the North Sea and Denmark. Shelveian structures near the Thor suture were incorporated in a poorly understood but complex, long-lived deformation zone that characterizes the TESZ. The TESZ separates the Baltic–East European craton from younger accreted lithosphere of western Europe over a strike length exceeding 2000 km. Accretion-related deformation in the TESZ possibly had already started during the Late Cambrian–Early Ordovician (Sandomierz orogenic phase of Belka et al., 2000) owing to accretion of a small peri-Gondwanan crustal block to Baltica. Evidence for Early to Late Ordovician orogenic events
(referred to as the Sardic or Eo-Variscan) is also preserved farther west in Europe (e.g.,
Armoric an terrane assemblage) from the Alps to Spain and Portugal (von Raumer, 1998; Handy
et al., 1999; Martínez Catalán et al., 2002; Matte, 2002; Robardet, 2002; and references therein).
They This evidence includes evidence for arc magmatism, ophiolite generation, polyphase
 deformation, and eclogite and high-grade metamorphism, which are generally thought to be
 related to formation of a short-lived volcanic-magmatic arc terrane(s) and backarc basin(s)
fringing the Eurasian margin (Fig. 1) of the Gondwanan supercontinent (Paleozoic Cadomia of
Stampfli et al., 2002) and an accretionary event. Ordovician orogenesis here does not represent a
major collisional event, but more of an amalgamation and/or reattachment of the peri-
Gondwanan, ribbon-like arc terrane(s) to the Eurasian margin of Gondwana. According to
Stampfli and Borel (2002), the reattached arc terranes were assembled into the long Hun
superterrane (Fig. 2), which became the leading edge of the Eurasian margin until it rifted off
Gondwana as a coherent ribbon continent during the latest Silurian and started to move north,
opening part of the Paleotethys in its wake. Reattachment of the Paleozoic Cadomian arc terrane
and closure of the intervening marginal basin during the Early to Middle Ordovician took place
after departure of Avalonia and concomitant eastward migration of the trailing Rheic Ocean
spreading center to the north of the Cadomian arc terrane. The eastward migration of the Rheic
spreading center may have put the arc terrane into compression (Stampfli and Borel, 2002) and
caused it to re-accrete. Final accretion of the Hun superterrane to Laurussia took place later
during the Variscan (Martínez Catalán et al., 1997, 2007; Matte, 2002).
The ribbon terranes fringing the Eurasian margin extended at least as far east as the
Himalayan margin of India (Stampfli and Borel, 2002; Cocks and Torsvik, 2002). The latter also
preserved evidence for extensive Cambrian–Ordovician orogenesis (e.g., Gehrels et al., 2006).
Accretion-reattachment of a peri-Gondwanan ribbon-like terrane to the Himalayan margin of
India, possibly kinematically related to coeval events along the European segment of this
Gondwana margin, is also an attractive mechanism for explaining Ordovician orogenesis here.

Central Asia Orogenic Belt (Altaids) and the Urals
The notion that the Ordovician globally was a time of development of a series of rapidly
evolving and amalgamating peri-cratonic arc terranes, which, when built on continental crust
commonly formed ribbon continents, may also be valid for the Central Asia orogenic belt of
central Asia and the Uralides. The central Asia orogenic belt and narrow Uralides formed mainly
during closing of the Aegir Ocean (Fig. 1) and separated the Siberian (Angara) craton from the
Baltic–East European craton to the east and the accreted terranes of the Tethysides to the south.
Ordovician orogenesis in the Central Asia orogenic belt mainly involved closure of marginal
basins and assembly of separated fragments of the ribbon-shaped Kipchak arc, according to
Şengör and Natal’in (1996). In addition, collision of part of the Kipchak arc with the Mugodzhar
arc-microcontinent, which was positioned outboard off the Baltic–East European craton’s eastern
margin and separated from it by the Sakmara-Magnitogorsk marginal basin (Fig. 1), may have
induced oroclastic bending and strike-slip imbrication of the Kipchak arc. The Mugodzhar arc-
microcontinent was later re-accreted to the East European craton during the middle–late
Paleozoic Variscan orogeny (Puchkov, 1997). Kröner et al. (2007), on the other hand, have
suggested that the central Asia orogenic belt is an accretionary orogen that formed by across-
strike, rather than by highly oblique accretionary processes, and the precursor of the Kipchak arc
was not a ribbon continent. In addition, detailed structural, petrologic, geochemical, and
geochronologic studies in the Ol’khon (Baikal) region in southern Siberia by Fedorovsky et al.
(2005 [[Q7: Fedorovsky is spelled Federovsky in the reference list. Q7]]) suggest that the
central Asia orogenic belt includes an Early Ordovician orogenic event. This event is represented by a polydeformed tectonic collage of ophiolite, arc volcanic, plutonic, and sedimentary rocks, produced during its accretion to the southern margin of the Siberian craton. Accretion was oblique, with strike-slip dominating over dip-slip tectonics. Metamorphic grade ranges from granulite facies against the Siberian craton, with little or no metamorphic overprint on the craton itself, to amphibolite facies to the south. Shortening is indicated by craton-vergent, ductile-brittle, Paleoozoic folds and thrusts in the adjacent Siberian craton, which formed during accretion of the tectonic collage to the craton. An intriguing element of this part of the orogen consists of linear to blocky bodies of marble associated with metasiliciclastic and volcanic rocks to the northwest that give way to dominance of less deformed and metamorphosed volcanic and plutonic rocks to the south.

**Famatinian Belt in South America**

Ordovician orogenesis was mainly localized along the proto-Andean margin, and is particularly well preserved in north-central Argentina, adjacent Chile and southern Peru, and Bolivia. This belt probably extends through Ecuador into Colombia. Orogenesis comprises subduction-related magmatism, metamorphism, deformation, and foreland basin development, which are variably referred to as the Famatinian and/or Ocloyic orogenic cycle (Ramos, 1988). During the early Paleozoic, the Famatinian belt was an active margin with development of a magmatic arc, which appears to have switched on several occasions from a compressional to an extensional state (e.g., Rapela et al., 1998). Transient compression of the magmatic arc caused localized deformation of arc plutons, which started during the latest Cambrian and/or earliest Ordovician (Pankhurst et al., 1998). Middle Ordovician compression of the Puná’s Famatinian arc-backarc basin system closed the backarc basin and led to concomitant deformation, retroarc–foreland basin development (Bahlburg and Hervé, 1997), and metamorphism in the underlying Mesoproterozoic crust of the Arequipa-Antefolla terrane (Loewy et al., 2004). Further south, orogenesis mainly involves accretion of the ribbon-like Precordillera-Cuyania microcontinent, whose provenance remains conjectural, although it is generally regarded as having been derived from Laurentia (Ramos et al., 1986; Astini et al., 1995; Thomas and Astini, 1996).

**Terra Australis Orogen in Australia and Antarctica**

Early Paleozoic orogenesis is also evident along the eastern margin of Australia, Tasmania, and the Transantarctic Mountains of Antarctica, which were positioned directly along strike of the proto-Andean margin of South America during that time. All three continents formed part of a very extensive convergent-accretionary margin that nearly circumnavigated the entire length of the Pacific margin of the Gondwana supercontinent (Terra Australis orogen of Cawood, 2005). The first phase of Ordovician orogenesis was represented by the waning stages of the predominantly Cambrian Ross-Delamerian orogen (Foden et al., 2006). However, age dating of the slaty cleavage of the most outboard Robertson Bay terrane (northern Victoria Land), suggests that deformation, at least locally, continued well into the Early Ordovician (Dallmeyer and Wright, 1992). This event was followed by the first phase of the Lachlan orogeny during the Late Ordovician (Foster et al., 1999; Foster and Gray, 2000; Collins, 2002). Both the Ross-Delamerian and Lachlan orogenies involve accretion of peri-cratonic volcanic arcs to the Australian-Antarctic cratons related to closure of the intervening marginal basins. The Ross-Delamerian orogeny may have involved more than one subduction zone (see reviews by Boger and Miller, 2004 [Q8: Boger and Miller, 2004: should this read Boger et al. to match the reference list? Q8]), and Cawood, 2005, and references therein) and was a tectonically complex event. Ophiolite obduction locally accompanied closure of the marginal basins. The
passive margin of the Lachlan backarc basin was built partly on the foundered Delamerian 368
Mountains and hence was filled by a large volume of clastic detritus from this orogen. This 369
clastic wedge was incorporated into a thick accretionary wedge during Late Ordovician–Silurian 370
backarc basin closure. The latter was achieved by a subduction of marginal basin crust beneath 371
both the passive and active sides of the basin. The Lachlan orogenic events continued into the 372
middle Paleozoic, around the Devonian-Carboniferous boundary, which led to full development 373
of the Lachlan segment (Foster and Gray, 2000) of the Terra Australis orogen.

DISCUSSION AND CONCLUSIONS

The evidence for worldwide Ordovician orogenesis clearly indicates that it was a global 376
phenomenon, but in contrast to the Proterozoic Grenvillian, late Paleozoic Appalachian, or 377
Tertiary Alpine-Himalayan, orogenic cycles, it did not involve true continent-continent 378
collision(s), and the major continents remained separated by wide oceans. Instead the Ordovician 379
is marked by formation of peri-cratonic arcs of great strike-length along the margins of 380
Laurentia, Baltic, Siberia, and Gondwana; linear orogenic belts, several characterized by low-381
grade metamorphism; marginal basins; suprasubduction ophiolites; ribbon microcontinents; 382
global high-water stands; and black shale deposition. Evidence for formation of true intra-383
oceanic juvenile arcs starved of continental-derived sediment (such as found in the Pacific today) 384
is rare or absent. The latter is significant because it indicates that subduction was localized along 385
continental margins. In addition, prior to the start of the Ordovician, during the Middle and Late 386
Cambrian (513–490 Ma), the margins of virtually all continental cratons underwent either the 387
onset of orogenesis or initiation of subduction. The latter commonly led to generation of large 388
tracts of suprasubduction zone ophiolite and associated boninite. For example, subduction had 389
initiated between 515 and 510 Ma directly outboard of the Ganderian, Baltic, and Laurentian 390
margins (van Staal et al., 1998), at a time when the Pacific margins of Antarctica and eastern 391
Australia saw the onset of Ross-Delamerian orogenesis. Early Cambrian Pampean orogenesis 392
along the West Gondwanan margin had started slightly earlier (ca. 525 Ma; Rapela et al., 1998), 393
but may also have been causally related. Combined, these phenomena suggest a causal link to a 394
global-scale plate reorganization. An attractive mechanism was the Early Cambrian terminal 395
suturing between East and West Gondwana (Boger and Miller, 2004 [[Q9: Boger and Miller, 397
2004: should this read Boger et al. to match the reference list? Q9]]; Cawood, 2005), which 398
undoubtedly must have had a major impact on the Early–Middle Cambrian global plate motion 399
budget. The rapid volume increase of oceanic spreading ridges that probably accompanied the 400
global-scale plate reorganization may also be responsible for the eustatic rise in sea level during 401
the Cambrian.

Surprisingly, subduction initiation in the Iapetus appears to have taken place near the 402
passive margins of Laurentia, Ganderia, and Baltica. In the Phanerozoic geological record there 403
are no obvious examples in which passive margins were directly transformed into active 404
margins. Margins are generally characterized by old, strong oceanic lithosphere and hence are 405
not favorable sites for subduction initiation (Cloetingh et al., 1982). The locations of nascent 406
subduction zones are generally thought to be controlled by the presence of a zone of weakness, 407
buoyancy, and compressive forces, the latter related to convergence (Mueller and Phillips, 1991, 408
and references therein). The presence of compressive forces is critical, because no combination 409
of a lithosphere-penetrating fault and negative buoyancy forces alone would produce a 410
subduction zone (Hall et al., 2003). For this reason, van Staal et al. (2007) proposed that 411
subduction had nucleated on old transform faults and/or abandoned spreading centers located 412
near the Laurentian margin of the Iapetus Ocean, created by an Early Cambrian inboard ridge

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jump(s), whereas on the opposite side of the Iapetus, near Avalonia and Ganderia, an extensive transform fault system had been created by late ridge-trench collisions during the late Neoproterozoic and Early Cambrian, respectively (Rogers et al., 2006). However, in light of the evidence discussed above and elsewhere (Buchan and Cawood, 2009), the possibility that subduction can initiate along a passive margin should not be ruled out. Erickson (1993) suggested that subduction could start near a passive margin, provided that the continent-oceanic interface becomes decoupled as a result of reactivation of old, rifting-related faults during strike-slip or extension.

Rollback of downgoing slabs probably was mainly responsible for the great tract of Early Ordovician peri-cratonic arcs and associated marginal basins that were present in the Iapetus, Aegir, and the Pacific Oceans. In particular, the propagation of the peri-cratonic arc-backarc systems that were present along the entire length of the Pacific and Iapetan margins of Gondwana (Australia, Antarctica, proto-Andean margin of South America, Ganderia, and Avalonia) and probably also Baltica was very impressive and comparable in scale to the active margins that surrounded the Pacific Ocean during the Mesozoic and Cenozoic. Tensional forces induced by slab rollback probably were also responsible for the terrane dispersal (Avalonia and Ganderia) along Gondwana’s Iapetan margin and their subsequent motion toward Laurentia (van Staal et al., 1998). Terrane dispersal along the Laurentian margin—e.g., Precordillera (Astini et al., 1995; Thomas and Astini, 1996), several internal southern Appalachian terranes (Merschat et al., 2005; Hatcher and Merschat, 2006), and Dashwoods (Waldron and van Staal, 2001; Cawood et al., 2001)—on the other hand was probably due to an inboard ridge jump from a spreading center that was positioned close to the Laurentian margin. It calved off ribbon-shaped slivers of continental crust from adjacent promontories (van Staal et al., 2007).

During the Ordovician, peri-cratonic terranes along the Gondwanan margin show several short-lived periods (10–20 Ma) during which large arc segments switched from extension to compression, commonly leading to closure of the trailing marginal basins or narrow seaways. The latter appears to be the main cause of Ordovician orogenesis along large segments of the Gondwanian margin (Penobscottian, Famatinian-Ocloyic, Sardin, late Ross-Delamerian, and early Lachlan) and probably also the margin of Baltica (Finnmarkian). Why the stresses along these margins switched from extension to transient compression is poorly constrained and not well understood at present. Collins (2002) proposed that compression was due to short-lived periods of flattening of the downgoing oceanic slab as a result of the entrance of relatively buoyant lithosphere, such as oceanic plateaus, into the subduction zone. Although this model is at first sight appealing and may be applicable to some events, it is difficult to accept that flattening of the circum-Gondwanan subduction zones more or less happened coevally (within a period of 20 m.y.) over a strike length exceeding 10,000 km simply because of arrival of buoyant oceanic crust; neither is there much evidence preserved for accretion of oceanic plateaus during the Early Ordovician along the Pacific and Iapetan margins of Gondwana. Perhaps the switch from extension to compression along the Pacific and Iapetan margins was due to far-field stresses generated by Gondwana’s drift and/or associated rotation during the Ordovician.

While the Ordovician mobile belts along the Gondwana margins were dominantly non-accretionary orogens sensu stricto (i.e., little or no accretion of truly exotic terranes, but rather involved re-accretion of rifted-off arc terranes: the extensional accretionary orogens of Collins, (2002), the Iapetan margin of Laurentia was the recipient of intra-Iapetan and/or Gondwana-derived arc terranes and/or microcontinents (van Staal et al., 1998; Hibbard, 2000; Hibbard et al., in press [[Q10: Anticipated year of publication? Q10]]; Merschat et al., 2005; Zagorevski et
al., 2006, in press) and hence, was steadily growing in size. This process continued throughout the Paleozoic and basically ended with the formation of Pangea during the Permian.

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Figure 1. Distribution of the main continents, oceans, subduction zones and arc-backarc systems, and microcontinents in the Early Ordovician (Tremadocian). The South Pole is situated in North Africa. The Pacific margin of Gondwana with the Ross-Delamerian belt of Australia and Antarctica that is continuous with the active proto-Andean margin, together forming a very long circum-Gondwanan subduction system, is just out of view. The Corel Draw figure originally made by Conall MacNiocaill for van Staal et al. (1998) was modified for this paper. Narrow volcanic arc terranes are indicated by ellipses. Dashed lines are used where the existence of subduction zones and/or transform faults is uncertain or contentious. Stippled structures may not have been active at this time. Arm—Armorica terrane; Aval—Avalonia; Boh—Bohemia terrane; Cadom—Paleozoic Cadomian arc terrane of Stampfli and Borel (2002); Car—Carolinia; Dashw and ND; arc—Dashwoods microcontinent and Notre Dame arc; Famat. arc—Famatinian arc-backarc system; Finnm arc—Finmark arc-backarc system; Gan—Ganderia; Høl.—Hølanda terrane; Ib—Iberia terrane; Kip arc—Kipchak arc; M—Meguma terrane; Mu arc—Mugodzhar arc-microcontinent; Pc—Precordillera terrane; Penobs. arc—Penobscot arc-backarc system; S-M—Sakmara-Magnitogorsk marginal basin; Svalb—estimated position of the Svalbard and Pearya terranes.

Figure 2. Paleogeography of the tectonic elements shown in Figure 1 in the Late Ordovician (450 Ma). Figure modified from van Staal et al. (1998). Abbreviations are the same as in Figure 1. Tac 2 and 3 refer to suture zones formed during the Taconian orogeny (van Staal et al., 2007). ACSW—Acadian seaway between Ganderia and Avalonia, the closure of which caused the Acadian orogeny; Hun—Hun superterrane of Stampfli et al. (2002); PVA—Popelogan-Victoria arc; RILA—Red Indian Lake arc; TEB—Tetagouche-Exploits backarc basin; Thor—Thor suture zone, forming part of the TESZ (Trans-European suture zone).