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The provenance of Jurassic and Lower Cretaceous clastic sediments offshore southwestern Nova Scotia

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Abstract: Jurassic and Cretaceous sandstones in the Shelburne subbasin and Fundy Basin, offshore Nova Scotia are poorly known, but are of current interest for petroleum exploration. The goal of this study is to determine the provenance of sandstones and shales, which will contribute to a better understanding of regional tectonics and paleogeography in the study area. Mineral and lithic clast chemistry was determined from samples from conventional cores and cuttings from exploration wells, using scanning electron microscope and electron microprobe. Whole-rock geochemical composition of shales was used to test the hypotheses regarding provenance of Mesozoic clastic sedimentary rocks in the SW Scotian Basin. Lower Jurassic clastic sedimentary rocks in the Fundy Basin contain magnetite, biotite and chlorite suggesting local supply from the North Mountain Basalt and Meguma Terrane, whereas pyrope and anthophyllite suggest small supply from distant sources. In the SW Scotian Basin, detrital minerals, lithic clasts and shale geochemistry from Middle Jurassic to Early Cretaceous indicate a predominant Meguma Terrane source, and transport by local rivers. Rare spinel and garnet grains of meta-ultramafic rocks, only in the Middle Jurassic at the Mohawk B-93 well, suggest minor supply from the rising Labrador rift, via the same river that transported distant sediments to the Fundy Basin. Lower Cretaceous sandstones from the Mohican I-100 well contain minor garnet, spinel and tourmaline from meta-ultramafic rocks, characteristic of sediment supplied to the central Scotian Basin at that time. The dominant Meguma Terrane provenance precludes thick deep-water sandstones in the eastern part of the Shelburne subbasin, but the evidence of Middle Jurassic distant river supply through the Fundy Basin is encouraging for deep water reservoir quality in the western part.

Keywords: Fundy Basin, Scotian Basin, sandstone, provenance, mineral chemistry, rivers, tectonics

Introduction

The determination of provenance for clastic sediments in sedimentary basins is a powerful indicator of quality and quantity of sand supply, especially frontier basins. In turn, this can provide useful information on clastic reservoir quality as well as constraints on regional tectonic evolution.

The Scotian Basin, offshore Nova Scotia, includes Mesozoic-Cenozoic sandstones and shales up to 15 km thick that host gas and lesser oil in the central part of the basin (Wade and MacLean 1990). The general characteristics of Mesozoic sediment sources for the central and eastern parts of the Scotian Basin are already known from previous studies (e.g. Tsikouras et al 2011; Pe-Piper and Piper 2012). However, little is known about the provenance of Middle Jurassic to Lower Cretaceous sedimentary rocks in the Shelburne subbasin, which represents the SW part of the Scotian Basin.

Recent hydrocarbon exploration has been of high interest in the deep-waters of the SW Scotian Basin, because potential Triassic–Jurassic source rocks are not overmature, as they are further to the east. Sea level low stands and hinterland tectonics may have created, at times, conditions favoring transport of turbidite sands from shelf-edge deltas into deep water reservoirs (Deptuck 2011; Deptuck and Campbell 2012). However, the only deep water well (Shelburne G-29) in this part of the basin is shaly; hence the reservoir potential of sandstones in the deep waters of the SW Scotian Basin has to be interpreted from the provenance of sandstones from wells on the shelf.

The overall goal of this study is to contribute to a better understanding of the petroleum geology in the SW Scotian Basin. The specific objectives are to determine the main rock sources for Middle Jurassic to Lower Cretaceous sedimentary rocks in the SW Scotian Basin and Lower

Jurassic sedimentary rocks in the Fundy Basin, as well as to predict possible patterns the rivers followed before entering and depositing in sedimentary basins.

Geological Setting and Stratigraphy

The Fundy Basin (Fig. 1), beneath the modern Bay of Fundy (Jansa and Wade 1975; Brown and Grantham 1992) is one of a series of half grabens that was initiated during the Triassic-Jurassic rifting of Pangaea (McIver 1972; Wade and MacLean 1990), and continued to develop as the North American plate drifted away from the African plate. Geologically, the basin formed at the boundary of the Avalon and the Meguma terranes of the Appalachians, which are separated by the Cobequid-Chedabucto Fault Zone (Wade et al. 1996). On land, to the SE, the basin is bounded by the latest Triassic North Mountain Basalt and the Meguma Terrane, whereas to the NW lie the Avalon and Gander terranes (Fig. 1).

The Meguma Terrane (Fig. 1) was deformed and assembled through the latest Proterozoic and Paleozoic and is the most outboard terrane of the Appalachian orogen, outcropping in southern Nova Scotia (Williams 1995). Neoproterozoic-Ordovician low grade metasedimentary rocks (slates, metapsammite and metapelite) up to 13 km thick, representing the Meguma Supergroup (White 2010) are intruded by Devonian granitoid plutons including the South Mountain Batholith (Clarke et al. 1997).

In Nova Scotia and southern New Brunswick, the Avalon Terrane is composed of Meso-Neoproterozoic sedimentary and volcanic rocks and overlying Cambrian-Ordovician shales and sandstones. These basement rocks are intruded by principally granitic plutons.

The oldest fill in the Fundy Basin consists of Upper Triassic Wolfville and Blomidon formations (Olsen et al. 1989; Schlische and Olsen 1990) that are capped by sub-aerial tholeiite

flows of the North Mountain Basalt at the Triassic—Jurassic boundary (Fig. 1). The Scots Bay Formation lies conformably above the North Mountain Basalt (Fig. 2) in the central part of the Fundy Basin. Due to erosion of almost 2 km of strata from the top of the succession, the age of the strata at the top of the formation cannot be determined, but likely extends to at least the Aalenian (Wade et al. 1996).

The Scotian Basin (Fig. 1) is considered a passive-margin sedimentary basin that was initiated around the same time as the Fundy Basin, under similar circumstances. The basin is separated in several depocenters, which from NE to SW, are the Laurentian, Abenaki (or Huron), Sable and Shelburne subbasins. To the northwest the basin is bounded by the Meguma Terrane, whereas the Atlantic Ocean represents the boundary to the southeast after oceanic spreading began in the mid Jurassic. Basement samples in offshore wells (e.g. Mohawk B-93 and Naskapi N-30) in the Scotian Basin, together with geophysical data, have shown that metasedimentary and igneous rocks of the Meguma Terrane comprise the basement beneath the present Scotian Shelf (Pe-Piper and Jansa 1999).

The subbasins are filled with Mesozoic-Cenozoic clastic sedimentary rocks, with up to 17 km of strata preserved beneath some parts of the Scotian Shelf. The oldest rocks are Upper Triassic–(?) Lower Jurassic evaporites of the Argo Formation and poorly sorted clastic sediments of the Eurydice Formation (Fig. 2), representing the initial phase of continental sedimentation, similar in composition and equivalent in age to the Upper Triassic Wolfville and Blomidon formations in the Fundy Basin. After the accumulation of the Argo and Eurydice formations, typical sabkha facies with shallow-water limestones and dolostones of the Iroquois Formation were deposited (McIver 1972) and are overlain by terrigenous clastic sediment of the Middle Jurassic Mohican Formation (Weston et al. 2012) in the eastern part of the SW Scotian Basin. In

the western part of the basin, coarse sandstones and shale representing the base of the Mohawk Formation were deposited. The Late Bathonian to Tithonian was a period of widespread carbonate deposition corresponding to the Abenaki Formation (Weston et al. 2012). The exception is in the western part of the SW Scotian basin, where from Oxfordian to Kimmeridgian sands of the Mohawk Formation continued to accumulate and the central part of the Scotian Basin, where from Kimmeridgian to Tithonian, sandy deltaic facies of the lower member of the Missisauga Formation, partly coeval with the upper strata of the Abenaki Formation in the west, prograded across the carbonate shelf and pass distally to the shales of the Verrill Canyon Formation (Cummings and Arnott 2005). By the Early Cretaceous, clastic sediments of the Missisauga and Logan Canyon formations extended across most of the shelf and fluvial equivalents, the Chaswood Formation, were accumulated and preserved in fault-bound basins on land. The outboard part of the SW shelf was the exception, as carbonates of the Roseway Formation equivalent in age to the Missisauga Formation continued to accumulate in areas far removed from clastic input (e.g. in the Mohawk B-93 well; Wade and MacLean 1990). The overlying Upper Cretaceous and Cenozoic successions consist mainly of shales and chalks.

Provenance of clastic sedimentary rocks in the Scotian Basin has been previously interpreted on the basis of several techniques, including chemical composition and modal abundance of detrital heavy minerals (e.g. Pe-Piper et al. 2009) and whole rock geochemical analysis (e.g. Pe-Piper et al. 2008, Zhang et al. 2014). Detrital heavy minerals are a powerful indicator of sediment provenance, but mineralogical techniques may not detect, for example, major igneous mafic sources that lack diagnostic stable detrital minerals, or sources which have been greatly diluted by minerals concentrated during polycyclic reworking (Zhang et al. 2014). Von Eynatten et al. (2003) pointed out that detrital heavy mineral analysis requires skill and is

time consuming. The use of whole-rock sediment geochemistry is challenging in areas lacking strongly contrasting source terranes, in areas where the terranes have parallel alignment to sedimentary basins, which is the case for the Scotian Basin, and also in rocks that have gained new elements and/or lost primary ones through diagenetic processes (Pe-Piper et al. 2008). However, geochemical studies may provide evidence for inputs from mafic sources that are poorly represented mineralogically (Ohta and Arai 2007) and also may indicate weathering conditions in the hinterland (Ruffell et al. 2002; Kahmann et al. 2008).

Methodology

Location and preparation of samples

Five wells (Figs. 1 and 3) penetrating Jurassic to Cretaceous rocks were selected for this study: Mohican I-100, Moheida P-15, Mohawk B-93 and Shelburne G-29 in the SW Scotian Basin, and Chinampas O-37 in the Fundy Basin. All the Middle Jurassic to Lower Cretaceous formations from the SW Scotian Basin were sampled in Mohican I-100, except the Mohawk, Shortland Shale and Verrill Canyon formations. In Moheida P-15 only the Roseway Equivalent, Abenaki and Iroquois formations were sampled, whereas in Shelburne G-29 the only formation sampled is Shortland Shale. The Lower Cretaceous Roseway Equivalent and Upper–Middle Jurassic Mohawk formations were sampled in Mohawk B-93. The Lower Jurassic Scots Bay Formation is the only interval sampled in Chinampas O-37 in the Fundy Basin.

Mohican I-100 and Moheida P-15 recovered 9 and 3 conventional cores, respectively (Fig. 3). A total of 28 rock slabs were cut from the back of conventional cores in these wells. In addition, a total of 53 cutting samples, with initial weight ~30 g, were obtained from intervals rich in sand in Mohican I-100 and Mohawk B-93.

Polished thin sections were made from 26 samples from cores in the Mohican I-100 and Moheida P-15 wells (Fig. 3). Fifteen samples (11 of which were shales) from Mohican I-100 contained sufficient material for whole-rock geochemical analysis. Clean samples were crushed in a shatterbox with an iron bowl. The rock powders were analyzed by Activation Labs, using their codes 4Lithoresearch and 4BI (Activation Labs 2012).

The samples were washed with liquid detergent and then with de-ionized water, before removing fine material using a 53 μm sieve and coarse material with a 250 μm sieve. The issue of whether to use a particular size fraction or a bulk sample is debated in the literature (Garzanti et al. 2009). Heavy minerals were separated using sodium polytungstate of 2.9 g/cm^3 density. Heavy minerals from adjacent samples were combined to obtain sufficient material for thin sections, resulting in 19 polished thin sections of heavy minerals from cuttings (Fig. 3). Details are provided in Dutuc (2015).

Analytical methods

Polished thin sections of both rock slabs and cuttings were studied first by a Nikon Eclipse E400 POL petrographic microscope equipped with a Pixel INK PL-A686C camera, to identify and determine distinctive groups of minerals and textures.

A LEO 1450 VP SME scanning electron microscope (SEM) was used to identify and determine the chemical composition of minerals in carbon coated polished thin sections by energy dispersive spectroscopy (EDS). The SEM has a maximum resolution of 3.5 nm at 30 kV, is equipped with an INCA Xmax 80 mm^2 silicon-drift detector (SDD) with detection limit of > 0.1% and uses a conventional high vacuum with a cooling system of liquid nitrogen to -180°C . A

copper standard was used for the calibration of the SEM. Back-scattered electron images (BSE) of sites comprising detrital minerals were taken for image analysis and textural interpretation.

Rutile was analyzed using a JEOL-8200 electron microprobe (EMP) by wavelength dispersive spectroscopy (WDS) with five wavelength spectrometers and a Noran 133 eV energy dispersion detector. The operating conditions were at 15 kV of accelerating voltage with a 20 nA beam current, a beam diameter of 1 μm and duration of analysis approximately 15 minutes. To avoid peak interference and for calibration of the EMP, samples of rutile were used as an internal standard.

The detrital heavy minerals (principally tourmaline, garnet, spinel and rutile) in this study have been identified and analyzed only in polished thin sections from cuttings. On the other hand, light detrital minerals (muscovite, biotite and chlorite) were analyzed only in polished thin sections of rock slabs from conventional cores.

Interpretation of sources from mineral chemistry and bulk rock chemistry

The chemical composition of most detrital minerals (heavy and light) were compared with a database built by Pe-Piper et al. (2009) and used by Tsikouras et al. (2011) to assess potential sources of Scotian Basin samples. For rutile we compared our data with that from Ledger (2013). For garnet geochemistry, we combined the southeastern Canadian database of Pe-Piper et al. (2009) with a set of 190 analyses from Deer et al. (1982) and distinguished 8 different garnet types (G1 to G8) with different sources.

The eight major elements that dominate source rock composition, together with a total of 26 trace elements (La, Ce, Nd, Sm, Eu, Gd, Yb, Lu, Y, Zr, Hf, V, Ni, Cr, Co, Sc, Nb, Ta, Th, Rb, Sr, Ba, Zn, Ga, U and Pb) were investigated in the analysed shales. Biplots of major elements, trace elements, together with REE plots and chondrite-normalized were produced using Minpet software. Most major elements are susceptible to post-depositional mobility and thus they are of limited value for provenance analysis (Taylor and McLennan 1985). Ti and Al, however, may be relatively immobile up to greenschist-grade metamorphic conditions (Pearce and Cann 1973). In addition, chemical weathering can influence the major-element geochemistry of sedimentary rocks, with the most significant changes resulting from alteration of feldspars and volcanic glass (Nesbitt and Young 1982). Trace elements in clastic deposits provide useful information about sedimentary provenance and composition of crustal source regions (Bhatia and Crook 1986). Trace elements including large-ion lithophile elements, high-field-strength elements, and rare earth elements are efficiently transferred into sedimentary rocks, are strongly excluded from seawater, and have low potential for post-depositional mobility (McLennan et al. 1993).

Results

Modal composition of detrital heavy mineral assemblages

The modal composition of detrital heavy mineral assemblages in the wells studied is presented in Table 1 and Figure 4. In most samples, the “heavy” separate ranged from 1.1% to 16.3% of the 53–250 μm cuttings. This fraction included inadequately separated light detrital and diagenetic minerals and abundant diagenetic heavy minerals, so that the detrital heavy minerals made up only between 0.7% and 21.5% of the “heavy” separate.

Mohawk B-93

Nine samples from Mohawk B-93 (Fig. 4), 3 from Middle Jurassic, 4 from Upper Jurassic and 2 from Lower Cretaceous formations, were studied for detrital heavy minerals. Ilmenite has been identified in all the formations and is the dominant detrital heavy mineral with percentage ranging from 23.8% to 85.5% in the Lower Cretaceous, with the exception of sample 1423.4 where zircon has a higher percentage (47.6%). In Middle and Upper Jurassic formations ilmenite abundance is greater than 60%. Tourmaline is also present in all the formations and is generally the second most abundant detrital heavy mineral with percentage that ranges from 4% in the Middle Jurassic to 19.4% in the Upper Jurassic. It tends to be less abundant in the Lower Cretaceous. Zircon, garnet, staurolite and apatite are the only other detrital heavy minerals present in all the formations sampled. Garnet and staurolite are most abundant (up to 13.9% and 6.9%, respectively) in the Middle Jurassic, and rare or absent in the Upper Jurassic and the Lower Cretaceous. Apatite is present in most samples studied, with percentage that ranges from 0.3% to 14.3% in the Lower Cretaceous. In the Middle and Upper Jurassic apatite abundance is not greater than 4.1%. Zircon is very variable in abundance at different stratigraphic levels (0.7% to 46.7%).

Other detrital heavy minerals like magnetite, monazite, rutile, chrome (Cr) spinel and xenotime are rare. Magnetite and rutile are both absent from the Upper Jurassic. Lower Cretaceous strata have 1% and 9.5% magnetite and ~2% rutile; Middle Jurassic strata have 2.4% magnetite and 0.7% rutile. Monazite (1.1%) was identified only in the Upper Jurassic, Cr-spinel (0.4%) in the Middle Jurassic and xenotime (0.3%) in the Lower Cretaceous.

Mohican I-100

Eight samples (Fig. 4) from Mohican I-100 have been studied for detrital heavy minerals: 3 from Middle Jurassic, 2 from Upper Jurassic and 3 from Lower Cretaceous formations. Ilmenite is the most abundant detrital heavy mineral in all the formations with weight percentage that ranges from 42.1% in Lower Cretaceous to 88.6% in Upper Jurassic strata. In Middle Jurassic formations ilmenite abundance is greater than 60%. Tourmaline is also present in all the formations studied, with abundance that varies from 6.9% to 42.1% in Lower Cretaceous strata. In Middle and Upper Jurassic tourmaline abundance is less than 21.7%. Zircon is the only other detrital heavy mineral identified in all the formations, ranging from 5.2% in the Lower Cretaceous to 17.6% in the Upper Jurassic. Garnet, generally, is sparse, ranging from 0.9% in the Upper Jurassic to 16.4% in the Middle Jurassic. The Lower Cretaceous has higher abundance in garnet (>5.2%) than Upper Jurassic (<2.4%). Staurolite percentage varies from 0.9% to 8.8% in the Lower Cretaceous. Middle Jurassic rocks do not have staurolite, whereas in the Upper Jurassic staurolite is rare (1.9%).

Other detrital heavy minerals such as andalusite, apatite, monazite, rutile and aluminum (Al) and chrome (Cr) spinel are rare. Andalusite (possibly including some sillimanite or kyanite) is present only in the Middle Jurassic and the Lower Cretaceous, having 2.4% and 0.9%, respectively. Apatite percentage ranges from 0.4% to 4.1% in the Middle Jurassic, whereas in other formations is rare (~1%). Middle Jurassic has monazite with abundance of 0.8% and 2.7%. Rutile, Al-spinel and Cr-spinel have been all identified only in the Lower Cretaceous, making up 5%, 2.6% and 0.9%.

Other wells

The dominant detrital heavy mineral at Chinampas O-37 is magnetite (96.7%), followed by rare amphibole (1.2%), apatite (0.9%), garnet (0.6%), ilmenite and tourmaline (each 0.3%). At Shelburne G-29 the dominant detrital heavy mineral is ilmenite (65.4%), followed by tourmaline (16%) and apatite (9.3%), with equal zircon and chrome spinel (4%) and rare aluminum spinel (1.3%). No heavy minerals were studied from Moheida P-15.

Chemical composition of detrital minerals (heavy and light) and lithic clasts

Tourmaline

Tourmaline was classified into 4 types using the discrimination diagram of Pe-Piper et al. (2009), where type 1 suggests a granitic rock source, type 2 a metapelitic or calc-silicate rock source, type 3 a meta-ultramafic rock source and type 4 a metapelitic-metapsammitic rock source. Only three of these types of tourmaline have been identified in the SW Scotian Basin: types 1, 3 and 4 (Figs. 5A, B, C), whereas no tourmaline was identified in the Fundy Basin.

In Mohawk B-93 well, type 4 is dominant and type 1 is present in some samples from each of the formations studied. Types 4 and 1 are also present in the Lower Cretaceous in Shelburne G-29. Type 4 tourmaline also predominates in the Mohican I-100 well. Type 1 is sub-ordinate and is present in all the formations sampled. Type 3 is restricted to the Middle Jurassic and the Lower Cretaceous. A few Type 1 tourmalines are also present in the Middle Jurassic of Mohican I-100.

Garnet

Type G1 garnet, almandine to spessartine from low grade metamorphic and felsic plutonic rocks, predominates in all wells at all stratigraphic levels (Fig. 6). One garnet in

Chinampas O-37 plots as 100% pyrope, classified as type G2 (ultramafic and metamafic rocks) (Fig. 6B) and another grain plots as type G1 (Fig. 6A). In Mohawk B-93, in addition to type G1, rare types G2 and G3 (high grade metamorphic rocks) are found principally in the Middle Jurassic. In Mohican I-100, Middle and Upper Jurassic formations have only type G1, whereas the Lower Cretaceous also includes some type G2.

Spinel

Spinel, typical of ophiolitic source rocks, is a rare detrital mineral in the Mesozoic sandstones in the SW Scotian Basin and was not found in the Fundy Basin. The only spinel found in Mohawk B-93 is Cr-spinel in the Mid Jurassic. In Mohican I-100, Cr- and Al-spinel are found only in the Lower Cretaceous. Rare Al- and Cr- spinel are found in the Lower Cretaceous in Shelburne G-29.

Rutile

Rutile is another rare detrital mineral. In Mohawk B-93, it is present only in the Mid Jurassic Mohawk Formation and the Lower Cretaceous Roseway Equivalent Formation, and in Mohican I-100 only from the Lower Cretaceous Upper Missisauga Formation. Individual grains from the Lower Cretaceous in both Mohawk and Mohican formations have chemistry similar to rutile from Appalachian metapelite and granite, whereas Mid Jurassic rutile in Mohawk B-93 includes these types plus rutile from granulite (see Figs. 4.40, 4.41 and 4.42 in Dutuc 2015).

Amphibole

Three different types of amphibole: tremolite, richterite-winchite and anthophyllite were chemically identified in the Lower Jurassic clastic sedimentary rocks in the Fundy Basin (see Fig. 5.24 in Dutuc 2015). However, total numbers are very small and the possibility of contamination from drilling mud cannot be excluded.

Biotite

Biotite is an abundant detrital mineral in both the SW Scotian Basin and the Fundy Basin. The chemical variation in biotite was used to determine potential rock sources, such as igneous, metamorphic and peraluminous granite after Fleet (2003) (Fig. 7A). For further discrimination of igneous biotite, the MgO-FeO-Al₂O₃ ternary plot of Abdel-Rahmen, (1994) showing fields for alkali, calcalkali and peraluminous rocks, was used (Fig. 7B).

Most biotite in the Lower Jurassic in Chinampas O-37 is metamorphic rather than igneous. The igneous biotite was sourced from calc-alkali and/or peraluminous rocks. Biotite in Mohawk B-93 is most common in the Upper Jurassic and has a similar range of chemistry and hence interpreted sources as in Chinampas O-37. Middle Jurassic formations in Mohican I-100 contain abundant biotite, whereas in the Upper Jurassic and the Lower Cretaceous biotite is rare. Biotite from the Middle Jurassic has an origin from both metamorphic and igneous rocks, whereas biotite analyzed from the Lower Cretaceous is entirely of igneous origin, sourced from calcalkali and/or peraluminous rocks.

Muscovite

Muscovite was analysed from polished thin sections of heavy mineral separates, and none suitable for analysis was found in Shelburne G-29 and Chinampas O-37. Mohawk B-93 has

muscovite in all cuttings samples studied. Chemically, Middle Jurassic muscovite identified in the Mohawk Formation is both metamorphic and igneous; Upper Jurassic in the same formation is mostly metamorphic, whereas all analyzed Lower Cretaceous muscovites identified in the Roseway Equivalent Formation are igneous (see Fig. 4.26 in Dutuc 2015).

Muscovite was analyzed from core samples in the Middle and Upper Jurassic in Mohican I-100 and Middle Jurassic in Moheida P15; all analyses plot in the field that represents metamorphic muscovite. Muscovite was also identified as inclusions in detrital quartz and detrital ilmenite in these wells.

Chlorite

Chlorite is abundant in the SW Scotian Basin and the Fundy Basin, but is rare in the one sample from Shelburne G-29. The FeO^t/MgO vs. $\text{SiO}_2/\text{Al}_2\text{O}_3$ discrimination diagram (Fig. 8) by Pe-Piper and Weir-Murphy (2008) was plotted with fields for diagenetic, metamorphic (detrital), and igneous (detrital) types of chlorite. To improve the diagram, which used chlorite analyses from Cretaceous rocks from the Orpheus graben, we added a new field that represents diagenetic chlorite based on chemical analyses obtained from diagenetic chlorite from sandstone samples of the Venture field (Gould 2007; Gould et al. 2010). Note that most analyses in the plot are by EMP/WDS, but SW Scotian Basin grains were analysed by SEM/EDS, resulting in slightly high estimates of Si and Fe (Sedge 2015).

In cuttings samples, almost half of the analyzed chlorite in the Lower Jurassic in Chinampas O-37 appears to be metamorphic in origin, whereas the other is diagenetic (Fig. 8D). For determination of diagenetic chlorite, additional morphological characteristics, such as shape and size were used (see Dutuc 2015). In Mohawk B-93, most of the chlorite plots in the

metamorphic field (Figs. 8A, C), with exception of a few grains in the Upper Jurassic that plot in the new diagenetic field (Fig. 8B). Lower Cretaceous chlorite in Shelburne G-29 is both detrital (metamorphic) and diagenetic.

In core and cutting samples, chlorite in Mohican I-100 can be observed in Mid Jurassic to Lower Cretaceous formations. Lower Cretaceous chlorites are concentrated only in the metamorphic (detrital) field (Fig.8A). On the other hand, chlorite from Upper Jurassic and Middle Jurassic sandstones tends to concentrate mostly in the detrital (metamorphic and igneous) fields with a few grains in the new diagenetic field (Figs. 8B, C).

Feldspar

Feldspar was systematically identified in polished thin sections and recorded as a light contaminant in heavy mineral separates (see Tables 4.3, 4.4, 4.5 and Fig. 4.25 in Dutuc 2015). Only albite, orthoclase and perthite were identified. K-feldspar is more abundant in the Lower Cretaceous at Mohican I-100 than in any other stratigraphic interval, whereas albite predominates in the Middle and Upper Jurassic at Mohawk B-93 (Figure 4.25 of Dutuc 2015).

Lithic clasts

Lithic clasts can only be identified in core samples from Mohican I-100 and Moheida P-15. Metamorphic and igneous lithic clasts were discriminated based on mineralogical composition and texture. The metamorphic lithic clasts (Figs. 9A, B) include metapsammite and metapelite with quartz + muscovite $\pm$ ilmenite $\pm$ chlorite. Some such lithic clasts show foliation, depending on grain orientation. The igneous lithic clast (Fig. 9C) is made up of quartz, muscovite, albite and K-feldspar, have granular texture, and are thus from granite. Metamorphic

clasts are present in Middle Jurassic and Lower Cretaceous strata in both wells, whereas igneous clasts were identified only in Middle Jurassic strata in Mohican I-100.

Whole-rock geochemistry of shales in Mohican I-100 well

Eleven shale samples from Mohican I-100 have 50%-70% SiO₂, <10% CaO, 6%-9% Fe₂O₃, 2.5%-6% K₂O and 8%-24% Al₂O₃ and are confirmed to be shale or mudstone using the criteria of Zhang et al. (2014). They are compared with one shale sample from Naskapi N-30 (Pe-Piper and Piper 2007).

Zr, Hf, Y and Yb

Zirconium (Zr) and Hafnium (Hf) are predominantly concentrated in ultra-stable zircon, together with Y and Yb which represent trace elements. There is a very good linear correlation between Zr and Hf (not illustrated) as well as between Zr and Y and Yb (Figs. 11A & 11B).

Ti, Cr, V, Nb, Zr and Ta

Titanium (Ti) is the main component in rutile and in general is mainly retained together with Al in clastic sedimentary rocks (Garcia et al. 1994). Element biplots are normalized to Al₂O₃ because some elements do show a strong correlation with abundance of clay minerals. In addition, Al is immobile and is less affected by diagenetic processes than other major elements such as Mg and Fe.

In the samples studied, there is no correlation between TiO₂ and Al₂O₃ (Fig. 10A), but discrimination between samples with different ages can be observed. Although samples from different stratigraphic layers tend to have similar concentration in Al₂O₃ they differ when it

comes to concentration in TiO_2 : Lower Cretaceous samples have the highest concentration, Upper Jurassic samples are subordinate, whereas Middle Jurassic samples have the lowest concentration.

$\text{Ti}/\text{Al}_2\text{O}_3$ tends to show a linear correlation with $\text{Zr}/\text{Al}_2\text{O}_3$ (Fig. 10B) and $\text{V}/\text{Al}_2\text{O}_3$ (not illustrated) for all stratigraphic levels, with the highest abundances in the Lower Cretaceous (Fig. 10B). $\text{Ti}/\text{Al}_2\text{O}_3$ (and also $\text{V}/\text{Al}_2\text{O}_3$ and $\text{Zr}/\text{Al}_2\text{O}_3$, not illustrated) show two different correlation trends with $\text{Cr}/\text{Al}_2\text{O}_3$ (Fig. 10C), $\text{Nb}/\text{Al}_2\text{O}_3$ (Fig. 10D) and $\text{Ta}/\text{Al}_2\text{O}_3$, one for the Middle Jurassic and another with higher $\text{Cr}/\text{Al}_2\text{O}_3$ and $\text{Ta}/\text{Al}_2\text{O}_3$ for the Upper Jurassic and Lower Cretaceous. Furthermore, Cr shows a good linear correlation with $\text{Ta}/\text{Al}_2\text{O}_3$ (Fig. 10E) and $\text{Nb}/\text{Al}_2\text{O}_3$ (not illustrated), but no correlation with Al_2O_3 (Fig. 10F). The one Lower Cretaceous sample from Naskapi N-30 tends to plot together with Middle Jurassic samples. In addition, Middle Jurassic samples show low Ti/Fe and Th/K ratios (Fig. 11E) compared to Upper Jurassic samples which have lower similar ratios than Lower Cretaceous samples.

Ce, La, Nd, Th and P

The rare earth elements Ce, La and Nd are the main components in monazite, whereas Th is usually trace element. Phosphorus (P) is often found in phosphate minerals such as monazite and xenotime, both previously identified in the Scotian Basin (Tsikouras et al. 2011; Li et al. 2012). The lack of correlation of Ce with Zr (Fig. 11C), but scattered positive correlation with P (Fig. 11D) suggests that Ce is concentrated mostly in monazite or xenotime rather than in zircon. The Ce vs Zr biplot suggests a low Ce/Zr ratio at lower stratigraphic levels, rather than higher, which is similar to that determined in one sample from the Lower Cretaceous Missisauga Formation at Naskapi N-30.

Rare Earth Elements (REE)

The stratigraphic variation in composition identified from element biplots between samples from Middle Jurassic, Upper Jurassic and Lower Cretaceous is also confirmed from plots (Fig. 12) of chondrite-normalised rare earth elements (REE).

Overall, REE tend to be more abundant in Lower Cretaceous and Middle Jurassic strata, with the exception 2 or 3 Middle Jurassic samples that overlap with Upper Jurassic samples. The Middle Jurassic samples show the most relative enrichment in the heavy (H) REE Yb and Lu and the most pronounced Eu anomaly. Upper Jurassic samples show the least relative enrichment in Gd.

Discussion

Sediment provenance

Lower Jurassic

Lower Jurassic rocks have been found only in the Fundy Basin and interpretation is limited by little available sample at the base of the Scots Bay Formation. Magnetite is the most abundant detrital heavy mineral (Table 1), but is otherwise generally rare in the Scotian Basin, making up 0.1% to 1% of the detrital heavy mineral separates (Tsikouras et al. 2011). It is locally abundant in the Middle Jurassic Mohican Formation in MicMac H-86 well (Fig. 13) (Li et al. 2012). Magnetite might be locally derived from the Late Triassic-Early Jurassic North Mountain Basalt (Olsen et al. 1989) (Fig. 1), or from minor Carboniferous and Lower Jurassic magnetite

mineralization along the surface trace of the Cobequid-Chedabucto Fault Zone (Fig. 1) (Ervine 1994; Murphy et al. 2011).

Metamorphic biotite and chlorite grains are abundant, and chemically similar to those in the Mid Jurassic of the SW Scotian Basin, which would be consistent with a local supply from higher grade metamorphic rocks of the southwestern Meguma Terrane (McKenzie and Clarke 1975; White 2010). The richterite-winchite sodic amphibole is known from the Cobequid Highlands to the northeast of the Fundy Basin (Papoutsis and Pe-Piper 2013), but other amphiboles are not diagnostic of source. Tremolite is known only from the Lower Cretaceous of the Musquodoboit E-23 well, where sediments were derived mostly from the Meguma Terrane (Pe-Piper et al. 2009). One garnet grain is 100% pyrope and another one is typical spessartine rich garnet of type G1 (Fig. 6), chemically similar to those from the Meguma Supergroup metasedimentary rocks (see Fig.5 in Pe-Piper et al. 2009). Overall, local supply of sediment is indicated by the available data, but the sources of single grains of tremolite, anthophyllite and pyrope are unknown.

Middle Jurassic

Middle Jurassic rocks include the Mohican and Mohawk formations and clastic intervals within the Iroquois and lower Abenaki formations (Fig. 4). Ilmenite is the dominant detrital heavy mineral, and is generally unaltered, suggesting a first-cycle source (Pe-Piper et al. 2005a). Common ilmenite with unoriented quartz, muscovite (Fig. 9C) and K-feldspar inclusions in the SW Scotian Basin show textures similar to those identified by Pe-Piper et al. (2005a, Figs.5c, 7c) inherited from igneous and metamorphic protoliths, most likely from the Meguma Terrane (Pe-Piper et al. 2004).

Tourmaline and garnet are common heavy minerals, but may be sourced from a wide range of rocks in southeastern Canada. However, the proportion of granitic, metamorphic and meta-ultramafic tourmaline in the SW Scotian Basin is similar to that in the Lower Cretaceous strata at the Naskapi N-30 well (Fig. 5D), where Reynolds et al. (2009) demonstrated from detrital muscovite and monazite geochronology that sandstones were exclusively sourced from the Meguma Terrane. Similar tourmaline varieties are found in the Lower Carboniferous Horton Group (Tsikouras et al. 2011), principally sourced from the Meguma Terrane (Murphy and Hamilton 2000).

The low number of grains of the ultra-resistant heavy minerals spinel, rutile and zircon (Table 1) identified in the SW Scotian Basin, compared to the abundance of ilmenite, suggests that the sediments are generally immature and sourced from crystalline rocks rather than older sedimentary rocks through recycling.

Further evidence of a Meguma Terrane source is provided by abundant greenschist grade metapelites and metapsammites as lithic clasts (Fig. 9) that resemble lithologies of the Meguma Supergroup metasedimentary rocks. In addition, muscovite inclusions (Fig. 9) in detrital quartz grains suggest that some muscovite was derived from peraluminous granites of the South Mountain Batholith (possibly reworked from the Horton Group). However, muscovite (see Fig. 4.26 in Dutuc 2015) and biotite chemistry (Fig. 7) indicates a dominant supply from metamorphic rocks and only minor supply from igneous rocks, including peraluminous granites in the case of igneous biotite (Fig. 7). Chlorite shows similar geochemistry (Fig. 8) to that from the Meguma Supergroup metasedimentary rocks.

Whole-rock geochemistry of Middle Jurassic shales at Mohican I-100 shows similar concentrations in trace elements (Zr, Nb, Cr, Ce and Ta) (Figs. 10B to 10E) and REE patterns

(Fig. 12) as one shale sample from the Lower Cretaceous Missisauga Formation at Naskapi N-30. As noted above for tourmaline, this interval in Naskapi N-30 shows exclusive derivation from the Meguma Terrane (Reynolds et al. 2009). The geochemical similarity implies that the Middle Jurassic at Mohican I-100 is sourced entirely from the Meguma Terrane.

In the Middle Jurassic at Mohawk B-93 some rare minerals suggest a small contribution from a source other than the Meguma Terrane. This evidence includes the presence of unusual type G2 garnet with chemical composition similar to those from metagabbro and anorthosite (Fig. 6), and from a few grains of spinel and type 3 tourmaline from meta-ultramafic rocks (Fig. 5). These mineral types are known from the central Scotian Basin, where detrital mineral geochronology (Pe-Piper and Piper 2012; Pe-Piper et al. 2014) suggests a source from ophiolites of western Newfoundland and metamafic rocks from the Grenville Province of Labrador.

Upper Jurassic

In general, the heavy mineral assemblage in the Upper Jurassic is similar to the Middle Jurassic, for example tourmaline (Fig. 5) and garnet (Fig. 6), and the same predominant Meguma Terrane source is inferred. Lithic clasts of Meguma Supergroup lithologies were not recognized in this interval, perhaps because available core samples from Mohican I-100 and Moheida P-15 are fine grained and distal, with interbedded limestones. However, the more proximal wells at Sambro I-29 and Naskapi N-30 (Fig. 13), where only cuttings are available, have thick sandstones which may include such lithic clasts.

Shale geochemistry at Mohican I-100 shows different behavior of Cr and Nb relative to Zr, REE and V in both Upper Jurassic and Lower Cretaceous samples compared with Mid Jurassic samples (Figs. 10B, C, D). Furthermore, Nb covaries with Cr (Figs. 10C, D) in the

Middle and Upper Jurassic and Lower Cretaceous and with Ta (Fig. 10E) only in the Upper Jurassic and Lower Cretaceous. High Nb and Ta values in Upper Jurassic shales from the central and eastern parts of the Scotian Basin were probably sourced from contemporary peralkaline volcanic ash transported by rivers draining Labrador (Zhang et al. 2014). The change in shale geochemistry in the Upper Jurassic is thus interpreted to be the distal influence of rivers supplying distant sediment to the central and eastern Scotian Basin.

Lower Cretaceous

Lower Cretaceous detrital petrology in the studied wells is similar to that in the Jurassic, indicating the continuing predominance of a Meguma Terrane source, particularly at Mohawk B-93. Lithic clasts of lithologies found in the Meguma Supergroup are again present at the Mohican I-100 well, perhaps indicating more vigorous uplift and erosion, or deposition in more proximal environments. Detrital biotite is predominantly of igneous origin (Fig. 7), in contrast to subequal igneous and metamorphic in the Upper Jurassic and predominant metamorphic in the Middle and Lower Jurassic. K-feldspar is also more abundant in the Lower Cretaceous at Mohican I-100 than in any other stratigraphic interval. The minerals tourmaline, garnet and muscovite show no such stratigraphic change in relative abundance of igneous and metamorphic varieties.

Detrital rutile was sufficiently abundant to be analyzed from Lower Cretaceous formations in Mohican I-100 and Mohawk B-93. Most grains plot in the field that represents metapelite on a Cr vs. Nb diagram (after Ledger 2013; see Fig.4.40 in Dutuc 2015). Nevertheless, a Zr vs Nb discrimination diagram (Ledger 2013) shows potential sources similar to those shown by tourmaline (types 1 and 4) and garnet (types 4 and 5), discussed previously in

the text, derived from an igneous (granite) and a metamorphic protolith (metapelite and metapsammite).

A few spinel grains were identified in Lower Cretaceous samples from the Mohican I-100 and Shelburne G-29 wells. Spinel represents a minor component in sandstones of the Chaswood Formation at Elmsvale Basin (Pe-Piper et al. 2004; Piper et al. 2008) and the Lower Carboniferous Horton Group in Nova Scotia (Murphy and Hamilton 2000; Tsikouras et al. 2011). Spinel is absent in Lower Cretaceous sandstones at Naskapi N-30 (Tsikouras et al. 2011) and in Chaswood Formation sandstones at Brierly Brook and Vinegar Hill (Pe-Piper et al. 2005b; Piper et al. 2007). Elsewhere in the Scotian Basin, especially in the central part, it is a prominent component of Lower Cretaceous rocks (Pe-Piper et al. 2009) derived from rock sources in the Labrador and Newfoundland (Fig. 1) through polycyclic reworking of Paleozoic sandstones, as indicated from the correlation with zircon abundance (Tsikouras et al. 2011). The absence of spinel in Naskapi N-30 confirms that it is not derived from the Meguma Terrane, and its absence in the Chaswood Formation at Vinegar Hill that it is not derived from the more inboard terranes of western New Brunswick. On the other hand, the small presence of spinel in the SW Scotian Basin has most likely similar source to that identified in other parts of the Scotian Basin. However, a potential source from the Lower Carboniferous Horton Group in Nova Scotia cannot be ruled out.

Other evidence of a source other than the Meguma Terrane is rare type 3 tourmaline (Fig. 5) and type G2 garnet, similar to that commonly found in Lower Cretaceous formations in the central and eastern Scotian Basin, likely sourced from metagabbros and anorthosites of the Grenville Province (Pe-Piper et al. 2009). Shale geochemistry shows similar trends to the Upper Jurassic, but with increased enrichment in elements such as Cr, Ta, Nb, Ti and Zr, probably

reflecting a higher proportion of reworking of polycyclic sediments. Some distal sedimentation from the Sable River supplying the central Scotian Basin reached the Mohican I-100 well site, but the predominant sediment supply remained from the Meguma Terrane.

Paleoclimate and river patterns

During the Late Triassic and Early Jurassic the Fundy Basin was located about 20° N of the equator and the climate was hot and semi-arid (Olsen and Et-Touhami 2008), confirmed by the presence of thick salt deposits of the Argo Formation (Fig. 2) on the Scotian Shelf (Wade and MacLean 1990). Sabkha-type anhydrite is present in the Middle Jurassic sandstones at Mohican I-100. Irregular but generally northward movement of North America through the Mesozoic (Beck and Housen 2003) meant that by Late Jurassic and Early Cretaceous, the Scotian Basin was located approximately 30°N of the equator (Irving et al. 1993), the Atlantic Ocean was approximately 1000 km wide (Ziegler 1989), and the climate was warm and temperate (Rees et al. 2000). Westerly equatorial currents were diverted into high northern and southern latitudes by the configuration of Pangaea, resulting in warm climate at mid latitudes (van Houten 1985). Humid conditions were recorded in the Chaswood Formation by the presence of cumulate ultisol and alfisol paleosols (Piper et al. 2009), although climate modelling (Hayward et al. 2004) suggests drier conditions inland with annual rainfall of ~1000 mm/a. The humid environment during the Early Cretaceous is also suggested by the high ratios of Ti/Fe and Th/K (Fig. 11E) which indicate leaching of K from illite which becomes kaolinite (see Fig.4.8A in Dutuc 2015) and alteration of ilmenite into leucoxene, pseudorutile and other alteration products (Pe-Piper et al. 2005a). Around the Jurassic-Cretaceous boundary, iron oolites are found in Moheida P-15

(Fig.4.15 in Dutuc 2015) and Dauntless D-35 in the eastern Scotian Basin (Weston et al. 2012), confirming a subtropical to tropical environment (Hallam and Bradshaw 1979).

In the Fundy Basin, paleocurrent studies by Leleu and Hartley (2010) suggest that Late Triassic clastic sediments were transported mostly by small local rivers draining areas of the Meguma and Avalon terranes and the North Mountain Basalt (Fig. 13). Despite extrusion of the North Mountain Basalt, this drainage pattern may have persisted into the Early Jurassic. A trunk river that ran westward along the Cobequid-Chedabucto Fault (CCFZ) would account for almost all the observed detrital petrology. The presence of single grains of pyrope and anthophyllite mean that the possibility of some long-distance supply (Fig. 13), draining inboard terranes of the Appalachians and/or the Canadian Shield, cannot be completely excluded.

Middle Jurassic to Lower Cretaceous strata at both the Mohican I-100 and Mohawk B-93 wells are dominated by Meguma Terrane supply. Small differences in modal abundances and chemical types of minerals suggests the SW Scotian Basin was supplied by more than one small river with local character draining areas of the Meguma Terrane throughout this time (Fig. 13). The presence of clinoforms extending from the Mohawk sands in the Mohawk B-93 well to the margin hinge zone suggests that during periods of low sea level sediments might have been transported and deposited in deep-water basins (Deptuck et al. 2011). In the Upper Jurassic, the Abenaki carbonate bank lay seaward of inner shelf sands at the Mohawk B-93 and Sambro I-29 wells (Fig. 13).

Minor more distant supply in the Mid Jurassic at Mohawk B-93 (type G3 garnet from the Grenville Province and spinel from ophiolites in Newfoundland), may have reached the Mohawk B-93 well through a river that entered the Fundy Basin through low lands from the northeast (Fig. 13). It is unlikely that such a river passed through the Orpheus graben and continued its

path along the CCFZ before entering the Fundy Basin, because such a path would have also supplied sediments to the Wyandot E-53 and Mic Mac H-86 wells (Fig. 13), where no distant mineral indicators are known (Li et al. 2012). The small number of distant detrital grains identified suggests that probably they reached the location of the Mohawk B-93 well by coastal longshore drift or shallow water tidal currents rather than direct river supply.

In the Lower Cretaceous, type G2 garnet, type 3 tourmaline and spinel at Mohican I-100 all suggest some contribution of sediment from the “Sable River” that drained from the rising Labrador rift through the Gulf of St. Lawrence to the central Scotian Basin (Zhang et al. 2014; Fig. 13). These more distant minerals were most likely deposited at Mohican I-100 as a result of occasional progradation of a delta distributary, or marine reworking of the Sable delta deposits to the west, where similar minerals derived from a river draining similar areas are abundant. The accumulation of limestones at Bonnet P-23 well and mostly limestone with sandy intervals at Mohawk B-93 (Fig. 4) shows that there was not a large input of sediment during the Early Cretaceous to the SW Scotian Basin.

The river that deposited the Chaswood Formation at Vinegar Hill, north of the Bay of Fundy, was different from that supplying the Chaswood basins of northern Nova Scotia, based on detrital monazite geochronology (Pe-Piper and MacKay 2006) and detrital mineralogy (Piper et al. 2007), both of which suggest sources in central and northern New Brunswick for the Vinegar Hill outcrops. This and other rivers likely have passed through low lands in the Fundy Basin, entered and finally deposited sediment in a “Shelburne delta” to the west of the Mohawk B-93 well (Fig. 13).

Tectonic implications

Local derivation of Early Jurassic sediment in the Fundy Basin suggests uplift of surrounding highlands, including the Meguma Terrane. The absence of rocks of definite Early Jurassic age in the Scotian Basin suggests uplift of the rift shoulder on what is now the Scotian Shelf adjacent to the basin depocentre beneath the present Scotian Slope (Fig. 13). Seismic profiles of the Scots Bay Formation (e.g. Figure 16 in Wade et al. 1996) show that it onlaps the North Mountain Basalt at the southern margin of the basin. This local uplift does not seem to have influenced the structural character of the Fundy Basin as sediments of the Scots Bay Formation continued to be deposited at least until Aalenian (Wade et al. 1996).

By the Middle Jurassic, peneplanation and subsidence of the seaward margin of the uplifted rift shoulder resulted in deposition of sediments of the Iroquois and Mohican formations were deposited in rift basins on the present Scotian Shelf (Fig. 13). The absence of clastic sediments at Bonnet P-23 (Fig. 4) implies that there was no major progradation of sediment supply through the Fundy Basin. However, the minor G2 garnet and spinel at Mohawk B-93 suggest some distant supply as discussed above. There is no evidence of supply of sediment from the North Mountain Basalt at that time, except the abundant magnetite in the Mic-Mac H-86 well (Li et al. 2012).

The timing of structural inversion of the Fundy Basin is unknown and no detrital evidence is known that might relate to major erosion of the Scots Bay Formation. Pe-Piper and Piper (2004) argued that it was the result of dextral strike slip movement on the Cobequid-Chedabucto Fault Zone that also produced the fault-bound basins of the Chaswood Formation in the Valanginian (Falcon-Lang et al. 2007; Pe-Piper and Piper 2012). On the other hand, uplift of the Labrador rift zone was initiated in the Kimmeridgian and rift-related tectonism began in the Tithonian in Newfoundland. In the eastern part of the Scotian Basin (Fig. 13), uplift of the inner

Scotian Shelf in the Kimmeridgian resulted in erosion and supply of Alleghanian muscovite to the Venture B-93 well (Reynolds et al. 2009, 2010). The regional near-base Cretaceous unconformity in the Bonnet P-23 well cuts out the Berriasian–Tithonian, but its relationship to the Fundy Basin is unclear. The greater abundance of sandy sediment supply and lithic clasts in the Lower Cretaceous may be a consequence of greater uplift of the Meguma Terrane at that time, which was also manifested by erosion of inner shelf Jurassic strata in the Valanginian–Hauterivian recorded in reworked palynomorphs in the Bonnet P-23 well (Weston et al. 2012).

Conclusions

1. Mineral abundance and chemistry in the Fundy Basin suggests that Lower Jurassic clastic sediments were derived mostly locally from the North Mountain Basalt and the adjacent Meguma and Avalon terranes. Rare pyrope and anthophyllite may indicate a minor contribution from a more distant source.
2. Middle Jurassic to Early Cretaceous clastic sedimentary rocks on the shelf in the Shelburne subbasin were derived predominantly from the Meguma Terrane. In the Middle Jurassic at Mohawk B-93 there was a small contribution from more distant source, indicated by type G2 garnet and spinel. In the Upper Jurassic–Lower Cretaceous of Mohican I-100 and Moheida P-15 there was a small contribution of sediment from the Sable River draining from the Labrador Rift through the Gulf of St Lawrence. This is indicated by mineral chemistry of some tourmaline and garnet grains, the presence of chrome spinel in sandstones, and the abundance of Cr, Ta, and Nb in shales.
3. Previous studies have shown that the Late Triassic to Middle Jurassic experienced a generally arid climate. Higher ratios of Ti/Fe and Th/K in Upper Jurassic shales, increasing further in

Lower Cretaceous samples suggest a transition to more humid conditions, also suggested by the deposition of iron oolites.

4. Uplift and erosion of the Meguma Terrane was most pronounced during the Middle Jurassic and Early Cretaceous, with increasing unroofing of granite. There is no clear evidence as to when tectonic inversion and erosion of the Fundy Basin took place.

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Table 1. Modal composition of detrital heavy mineral assemblages**Figure Caption**

Fig. 1. Regional geological map of Eastern Canada showing bedrock geology and location of the five named wells investigated in detail in this study. Other wells mentioned in text are: B-93 = Venture B-93; E-23 = Musquodoboit E-23; H-86 = MicMac H-86; I-29 = Sambro I-29; N-30 = Naskapi N-30; P-23 = Bonnet P-23; E-53 = Wyandot E-53. SMB = South Mountain Batholith, NMB=North Mountain Basalt. (Map modified from Sawatzky and Pe-Piper 2013 and Williams and Grant, 1998).

Fig. 2. Lithostratigraphy of Mesozoic formations in the SW Scotian Basin and the Fundy Basin. Lithostratigraphy for the SW Scotian Basin is modified from Weston et al. (2012) using information from Wade and MacLean (1993), whereas for the Fundy Basin is modified from Leleu and Hartley (2010).

Fig. 3. Schematic stratigraphic columns for the wells studied, showing type, age and stratigraphic location of analyzed samples. Biostratigraphy and lithostratigraphic picks are taken from figure 4.

Fig. 4. Modal abundance of detrital heavy minerals in cutting samples and biostratigraphic correlation. Lithology is based on cuttings with data taken from well reports at the Canada Nova Scotia Offshore Petroleum Board (CNSOPB). Biostratigraphy and lithostratigraphic picks are based on MacLean and Wade (1993), Weston et al. (2012) and OETR (2011).

Fig. 5. (A-C) Plot of analyzed tourmaline grains from the SW Scotian Basin on the classification diagram of Pe-Piper et al. (2009), modified after Henry and Guidotti (1985). (D) Cumulative relative abundance of tourmaline types acquired by using SEM analyses (cf. Li et al. 2012). Naskapi from Pe-Piper and Piper (2007).

Fig. 6. Chemical variation in garnets from known crystalline rock sources, the SW Scotian Basin and the Fundy Basin. Garnet chemistry from potential rock sources from Deer et al. 1982 (black dots) and Pe-Piper et al. 2009 (red dots) was used to identify eight provenance types (G1 to G8). Three of these provenance types are equivalent to the types identified by Pe-Piper et al. (2009): G1 = 4 and 5, G2 = 1B and 3, and G3 = 1A.

Fig. 7. (A) Chemical variations in biotite showing potential rock type sources. The fields for potential rock sources are taken from Fleet (2003). (B) Fields showing chemical discrimination of igneous biotite based on data taken from Abdel-Rahmen (1994), where A = alkali, C = calcalkali, P = peraluminous.

Fig. 8. Chlorite discrimination diagram from Weir-Murphy (2004) for the SW Scotian Basin and the Fundy Basin. The discrimination fields were plotted using chlorite analyses from the Cretaceous rocks from the Orpheus graben (Pe-Piper and Weir-Murphy 2008). The diagenetic chlorite field that covers a small area in the metamorphic field is after Sedge (2015). Data for detrital chlorite from the Meguma Supergroup metasedimentary rocks (small black dots) is provided by Dr. Chris White (pers. comm.)

Fig. 9. SEM back-scattered electron images of sedimentary rocks showing texture and mineralogical composition of: A) greenschist metamorphic lithic clast in the Mid Jurassic in Moheida P-15 showing characteristic foliation; detrital quartz with muscovite inclusion. Lithic clasts in Mohican I-100 are: B) metapelite or metapsammite in the Mid Jurassic, C) metapelite or metapsammite (M) and granite (G) in the Mid Jurassic together with ilmenite with muscovite and quartz inclusions, and D) greenschist metamorphic lithic clast, metapsammite or metapelite together with detrital quartz with muscovite inclusion. (qz=quartz, ms=muscovite, ilm=ilmenite, bt=biotite, chl=chlorite, kfs=K-feldspar, ab=albite, tur=tourmaline)

Fig. 10. Variations of TiO_2 with Al_2O_3 (A), Ti with Zr, Cr and Nb (B, C & D), and Cr with Ta and Al_2O_3 (E & F) for shale samples.

Fig. 11. Variations of Zr with Yb and Y (A & B), Ce with P and Zr (C & D) and Ti with Th (E) for shale samples.

Fig. 12. Rare earth element diagrams for (A) Upper Jurassic and Lower Cretaceous shale samples and (B) Mid Jurassic and Upper Jurassic shale samples.

Fig. 13. Schematic map showing potential rivers and sources for Lower Jurassic clastic sediments in the Fundy Basin (A) and Mid Jurassic to Lower Cretaceous clastic sediments in the SW Scotian Basin (B, C, & D). The red dots on land represent the Lower Cretaceous Chaswood Formation and the blue dashed lines represent possible rivers proposed by Pe-Piper and Piper (2012).

Table 1. Modal abundance of detrital heavy mineral assemblages

Well	Chinampas O-37	Mohawk B-93								Mohican I-100								Shelburne G-29	
Age	EJ	MJ			LJ				EK		MJ			LJ		EK			EK
Stratigraphic Unit	SB Fm	MK Fm							RE Fm		IR Fm		AB Fm	RE Fm		MM Fm	UM Fm	LC Fm	SS Fm
Depth (m)	301.75	2058.92	1993.41	1932.43	1892.8	1787.64	1743.45	1650.48	1577.33	1423.4	4206.24	3852.67	3474.72	2685.28	2584.7	2389.63	2203.7	1798.32	3635
wt% "heavies":	10.82	5.72	3.18	3.14	1.71	6.55	8	2.78	2.71	16.28	5.15	3.42	1.11	3.72	4.08	4.81	3.6	8.96	4.83
% detrital heavy minerals in "heavies"	8.42	5.21	5.33	13.06	16.23	5.94	3.95	21.46	5.92	1.34	2.33	12.85	10.3	4.15	4.64	0.74	3.67	1.44	2.14
Amphibole	1.2	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Andalusite	—	—	—	—	—	—	—	—	—	—	—	0.4	—	—	—	—	—	2.9	—
Apatite	0.9	2.4	4.1	0.7	0.5	—	—	0.3	1.0	14.3	—	0.4	4.1	1.2	—	—	0.9	—	9.3
Garnet	0.6	1.6	13.9	5.5	2.2	—	1.4	1.7	3.0	—	—	16.4	1.4	2.4	0.9	5.2	7.8	—	—
Ilmenite	0.3	81.5	59.8	81.2	76.8	72.3	80.8	85.5	68.0	23.8	81.0	59.9	60.9	68.2	88.6	42.1	62.9	73.6	65.4
Magnetite	96.7	2.4	—	—	—	—	—	—	1.0	9.5	—	—	—	—	—	—	—	—	—
Monazite	—	—	—	—	1.1	—	—	—	—	—	—	0.8	2.7	—	—	—	—	—	—
Rutile	—	—	—	0.7	—	—	—	1.0	2.0	—	—	—	—	—	—	—	5.0	—	—
Aluminum Spinel	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	2.6	—	1.3
Chrome Spinel	—	—	—	0.4	—	—	—	—	—	—	—	—	—	—	—	—	0.9	—	4.0
Staurolite	—	4.8	4.1	3.0	—	6.9	5.5	1.7	3.0	—	—	1.9	—	1.2	—	5.2	0.9	8.8	—
Tourmaline	0.3	4.0	13.9	5.9	19.4	6.9	11.0	8.8	7.0	4.8	9.5	13.7	21.7	9.4	7.0	42.1	6.9	8.8	16.0
Xenotime	—	—	—	—	—	—	—	0.3	—	—	—	—	—	—	—	—	—	—	—
Zircon	—	3.2	5.7	2.6	—	13.9	1.4	0.7	15.0	47.6	9.5	6.5	9.5	17.6	3.5	5.2	12.1	5.9	4.0
Total no of grains (heavy only)	345	124	122	270	181	101	73	299	100	21	21	262	74	85	115	19	116	34	75
Total (heavy+light+diagenetic)	4096	2380	2285	2067	1115	1700	1845	1393	1688	1563	1987	2038	718	2044	2478	2535	3195	2358	3497

Note: "Heavies" include detrital heavy and light minerals, as well as diagenetic minerals. Wt% of "heavies" is percent of total sample with grains >250 µm and <53 µm. % detrital heavy minerals "heavies" is based on grain counts (both chemical analyses and BSEI) and represents the total counts of detrital heavy minerals within the total count of grains (heavy+light+diagenetic) in the whole sample. The percentage of each individual mineral represents the total number of grains counted from EDS chemical analyses and BSEI within the total number of grains (heavy only). no=number, EDS=Energy Dispersive Spectroscopy, BSEI=Backscattered Electron Image, E=Early, M=Middle, L=Late, J=Jurassic, K=Cretaceous, SB=Scots Bay, MK=Mohawk, RE=Roseway Equivalent, IR=Iroquois, AB=Abenaki, MM=Middle Mississauga, UM=Upper Mississauga, LC=Logan Canyon, SS=Shortland Shale, Fm=Formation

Table 1S. Whole-rock geochemical analyses of mudstones from Mohican I-100 and Naskapi I

Well					
Formation	Roseway Equivalent				Aber
Depth (m)	2532.73	2533.49	2539.05	2539.18	2840.49
Lithology	mudstone	mudstone	mudstone	mudstone	mudstone
Major element (wt%)					
SiO ₂	51.71	50.30	52.47	55.70	49.66
TiO ₂	1.11	1.05	1.13	1.10	0.85
Al ₂ O ₃	12.90	12.41	17.92	14.90	15.00
Fe ₂ O ₃ T	6.66	7.40	6.34	5.99	5.43
MnO	0.04	0.04	0.03	0.03	0.05
MgO	2.62	2.58	2.39	2.19	1.76
CaO	7.86	7.99	4.25	3.44	8.89
Na ₂ O	0.65	0.54	0.70	0.57	0.76
K ₂ O	2.66	2.29	3.02	2.47	2.23
P ₂ O ₅	0.16	0.20	0.14	0.20	0.05
LOI	12.28	13.52	12.08	11.34	13.72
Total	98.65	98.31	100.50	97.93	98.39
Trace element (ppm)					
Sc	14	13	18	15	15
Be	3	3	3	3	3
V	162	153	172	155	125
Cr	110	170	130	150	120
Co	20	17	24	18	19
Ni	56	40	74	50	61
Cu	22	30	25	20	23
Zn	67	80	94	90	57
Cd	0.5	-	0.5	-	0.5
S	0.982	-	0.869	-	0.816
Ga	18	17	24	21	21
Ge	1.7	1.5	1.9	1.6	2
As	19	14	13	11	7
Rb	97	99	118	108	130
Sr	300	266	222	181	441
Y	23	22.7	25.7	25.4	21.9
Zr	268	204	226	220	204
Nb	40.5	39.7	33	37.2	23.5
Mo	2	2	2	2	2
Ag	0.3	0.5	0.3	0.7	0.3
In	0.1	0.1	0.1	0.1	0.1
Sn	2	2	3	2	2
Sb	0.3	0.2	0.4	0.2	0.2

Cs	6	5.4	7.5	6.1	8.8
Ba	231	212	295	250	237
La	47.3	42.1	47.6	45.7	34.9
Ce	98.5	76.9	102	89.4	73.8
Pr	10	8.79	10.5	9.93	7.67
Nd	34.4	31.4	36.8	35.7	26.7
Sm	5.8	5.5	6.75	6.74	4.54
Eu	1.28	1.3	1.41	1.5	0.962
Gd	3.89	4.54	4.62	5.21	3.41
Tb	0.69	0.73	0.75	0.85	0.57
Dy	4.2	4.25	4.71	4.86	3.79
Ho	0.86	0.84	0.97	0.96	0.78
Er	2.41	2.45	2.65	2.75	2.22
Tm	0.36	0.377	0.4	0.417	0.332
Yb	2.34	2.4	2.71	2.67	2.36
Lu	0.393	0.334	0.451	0.395	0.398
Hf	5.9	5.4	4.9	6	4.8
Ta	1.96	2.15	1.76	2.07	1.46
W	1.1	1.3	1.4	1.5	1.4
Tl	0.4	0.27	0.5	0.37	0.51
Pb	4	9	9	10	14
Bi	0.1	0.1	0.1	0.1	0.1
Th	10.3	9.12	12.5	11.6	10.8
U	3.06	2.82	2.88	3.16	2.2

N-30 wells

Mohican I-100					
Naki	Mohican		Iroquois		
2842.05 mudstone	3696.55 mudstone	3697.95 mudstone	3960.68 mudstone	4332.43 mudstone	4336.54 mudstone
51.52	43.02	62.26	41.47	45.64	56.19
0.98	0.63	1.07	0.65	0.69	0.79
19.28	18.60	8.14	13.62	12.90	11.95
6.31	6.34	1.60	5.06	6.74	6.13
0.04	0.06	0.01	0.05	0.29	0.08
1.90	4.59	0.60	7.04	5.71	5.17
2.14	6.17	9.85	8.91	5.52	2.68
0.82	0.84	1.33	0.89	0.39	0.27
2.98	4.50	1.98	3.56	4.96	4.36
0.05	0.08	0.07	0.05	0.11	0.12
14.76	12.84	6.74	17.60	16.44	10.98
100.80	97.67	93.65	98.90	99.40	98.71
20	16	5	12	14	12
3	3	1	2	2	2
159	111	41	107	117	105
140	100	40	80	80	80
22	24	5	22	23	26
80	53	12	37	38	39
33	43	39	29	7	6
68	976	21	48	57	58
0.5	9.2	0.5	0.5	0.5	0.5
0.968	1.86	5.62	1.55	1.47	1.07
25	21	10	18	19	15
1.9	1.8	1.5	1.6	2	1.9
5	19	5	16	10	8
169	167	68	114	131	102
189	160	904	964	307	248
20.2	15.8	15.7	12.1	23.8	27.9
182	142	245	191	170	324
27.4	11.4	10.4	12	12.6	12.7
2	2	2	10	2	2
0.4	0.3	0.3	0.5	0.3	0.3
0.1	0.1	0.1	0.1	0.1	0.1
2	3	4	2	2	2
0.3	1.6	0.3	0.7	1.2	1

11.4	11.9	3.1	6.1	14.8	9.8
265	549	253	465	323	358
41.6	31	31.7	22.3	38.4	27.5
86.5	62.4	73.8	43.9	83	60.9
8.42	6.48	7.95	4.25	8.91	6.55
27	23.1	30.6	14.7	33.7	24.6
4.17	3.68	5.83	2.59	5.91	5.27
0.825	0.767	1.05	0.493	1.24	1.11
2.86	3.2	3.63	1.83	4.26	4.71
0.53	0.51	0.49	0.29	0.64	0.77
3.61	3.05	2.83	1.91	4.13	4.82
0.76	0.63	0.57	0.43	0.83	0.95
2.32	1.93	1.54	1.33	2.31	2.71
0.358	0.299	0.219	0.218	0.338	0.4
2.53	2.08	1.5	1.56	2.29	2.74
0.415	0.346	0.233	0.262	0.387	0.459
4.2	3.7	5.2	4.3	4.1	7.3
1.71	0.83	0.86	0.8	0.82	0.86
1.3	1.6	1.2	1.3	0.7	1.1
0.63	0.66	0.33	0.91	0.37	0.33
11	62	3	24	12	11
0.1	0.1	0.1	0.1	0.2	0.1
12.5	12	6.99	9.25	9.67	9.47
2.34	2.56	1.52	5.57	2.69	2.82

Naskapi N-30
Missisauga
1475.53 mudstone
58.85
0.81
19.27
3.52
0.01
0.32
0.08
0.52
2.24
0.08
12.73
98.43
15
4
95
80
34
72
35
138
0.5
2.33
26
6.2
16
102
94
28
269
18.3
2
0.3
0.1
4
0.3

Draft

9.6
490
43.8
87.7
10
36.6
7.18
1.45
5.95
0.97
5.36
-
2.83
0.429
2.83
0.439
7.5
1.41
0.5
2.24
39
0.1
12.2
3.74

Draft

Dutuc et al. Fig.1

Dutuc et al. Fig.2

Dutuc et al. Fig.3

Dutuc et al. Fig.4

Dutuc et al. Fig.5

Dutuc et al. Fig.6

Dutuc et al. Fig.7

Dutuc et al. Fig.8

Dutuc et al. Fig.9

Dutuc et al. Fig.10

Dutuc et al. Fig.11

Dutuc et al. Fig.12

Dutuc et al. Fig.13