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Seismic and gravity constraints on the crustal architecture of the Intermontane terranes, central Yukon

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Draft

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A.J. Calvert, N. Hayward, R. Vayavur, and M. Colpron. Seismic and gravity constraints on the crustal architecture of the Intermontane terranes, central Yukon

Abstract

In 2004, two seismic reflection lines were shot across the Mesozoic Whitehorse trough and adjacent terranes. 3-D first arrival tomographic inversion is used to constrain lithology to 800-1200 m depth, and surface structures are extrapolated into the middle crust using the coincident reflection data. In the Yukon-Tanana terrane, the metasedimentary Snowcap assemblage is characterized by velocities of 4.5-5.5 km/s, while in Quesnellia, velocities of 5.0-6.0 km/s occur at 500 m depth, and probably represent igneous rocks of the Tatchun batholith. Across the Whitehorse trough, velocities >4.0 km/s correspond to clastic rocks of the Jurassic Laberge and Triassic Lewes River groups; velocities <4.0 km/s probably present the clastic Jurassic to Cretaceous Tantalus Formation. Several near-surface units with velocities of 2.0-3.0 km/s are identified; some correlate well with volcanic rocks of the Upper Cretaceous Carmacks Group, but others could be attributable to alluvial deposits or faulting. The Big Salmon fault is interpreted to dip southwest, implying that rocks of the Yukon-Tanana terrane extend beneath Quesnellia. Stikinia and Quesnellia underlie up to 5-8 km of Triassic to Early Cretaceous sedimentary strata, and appear to be a single allochthon within an 18-20 km deep synform above the Yukon-Tanana terrane, which we name the Northern Intermontane synform. In general, reflection geometries in the upper crust are complex, but are consistent with large-scale imbricate structures that have been dissected into numerous blocks by displacement along moderately to steeply dipping strike-slip faults, which may be part of a crustal-scale flower structure extending to the base of the crust.

Keywords: Whitehorse trough, seismic reflection, gravity inversion, Stikinia, Quesnellia, Yukon-Tanana

Introduction

The Whitehorse trough is the northernmost of the Intermontane sedimentary basins of the Canadian Cordillera, which lie between the Pelly Mountains or Rocky Mountains in the east and the Coast Mountains in the west (Mossop et al. 2004). The Whitehorse trough is an elongate Mesozoic basin that extends over 650 km northwestward from northern British Columbia to just north of Carmacks in south-central Yukon (Fig. 1). The basin is estimated to contain $82 \times 10^9 \text{ m}^3$ (2920 BCF) of gas and $17 \times 10^6 \text{ m}^3$ (107 MMBO) of oil (Hayes 2012), but with large uncertainties due to limited available structural, stratigraphic and thermal maturity information (English et al. 2005, Lowey et al. 2009). The basin remains largely unexplored, and no wells have been drilled. In 2004, 170 km of regional seismic reflection survey were acquired along existing roads across the northern Whitehorse trough near Carmacks (Fig. 2) to better understand its structural and stratigraphic setting (White et al. 2006). An interpretation of the upper 5 km using the two seismic lines together with a summary of the acquisition and processing were presented by White et al. (2012), who inferred that as much as 3000 m of Jurassic strata of the Laberge Group were deposited in the Whitehorse trough, but many near-surface structures remain poorly defined. Although the crust-mantle boundary can be identified below some parts of these seismic lines at approximately 35 km depth, the structural relations between the sedimentary basin, the underlying crystalline crust, and the adjacent terranes is uncertain due to variable data quality.

To provide additional constraints on structural and lithological heterogeneity up to a depth of 1.5 km, we derive a P-wave velocity model using 3-D tomographic inversion of the first arrivals recorded in the reflection survey, and then correlate velocity changes with faults inferred from discontinuities in seismic reflection images. By reprocessing the reflection data to remove

interfering coherent noise, we are able to interpret faults that bound the Whitehorse trough and the major terranes. We identify the continuation of some faults to approximately 20 km depth in the middle crust, and define a broad 18-20 km deep synform that contains the Stikinia and Quesnellia terranes, which underlie much of the intermontane region of the Canadian Cordillera. Both 3-D inversion of gravity data around the seismic lines and the seismic interpretation are broadly consistent with Triassic to Early Cretaceous sedimentary rocks extending to 5-8 km depth within this broader structure. From the geometry of the terrane bounding faults, we suggest that the Yukon-Tanana terrane here underlies both Stikinia and Quesnellia, constraining the process of terrane accretion in this part of the Cordillera.

Geological setting of the Whitehorse trough

The Whitehorse trough is underlain by the Devonian to Middle Jurassic Stikinia and Quesnellia arc terranes, which were accreted to North America by at latest the Middle Jurassic; the two terranes are believed to be closely related having formed in a similar tectonic setting proximal to the continent (Nelson et al. 2013). In central Yukon, Quesnellia and Stikinia are in contact, partly along the Teslin fault, but in southern Yukon and northern British Columbia the two terranes are separated by the Cache Creek terrane accretionary complex, which contains exotic elements. Quesnellia and Stikinia are themselves bounded to the east, west and north by the Devonian to Permian Yukon-Tanana terrane, which developed as a continental arc that was separated from North America by a back arc basin, fragments of which are now preserved as the Slide Mountain terrane east of the Yukon-Tanana terrane. At the latitude of the Carmacks seismic transect, the Yukon-Tanana terrane is separated from the Slide Mountain terrane by the Tummel fault, and the Cassiar terrane of North American affinity lies east of the Slide Mountain terrane (Colpron et al. 2005). The Yukon-Tanana terrane accreted to North America during and after the Late Permian

(Beranek and Mortensen 2011). The terrane is interpreted to form, in part, the basement to Quesnellia and Stikinia, with its present occurrence outboard of Stikinia possibly the result of oroclinal bending that also trapped the Cache Creek terrane between Stikinia and Quesnellia (Mihalnyuk et al. 1994; Nelson et al. 2013). The Whitehorse trough comprises Early to Middle Jurassic sedimentary strata that were deposited in a forearc setting, and subsequently shortened as the basin evolved into a synorogenic piggyback basin at the start of the Cordilleran orogeny (White et al. 2012; Colpron et al. 2015, and references therein).

Stratigraphy

Near the northern Whitehorse trough, Stikinia rocks include the Paleozoic Takhini assemblage and the overlying Upper Triassic Lewes River Group (Figs. 2 and 3). The Takhini assemblage comprises metamorphosed volcanic, volcanoclastic and carbonate rocks, which are exposed on the western edge of the basin (Hart 1997). The overlying Lewes River Group outcrops in and around the Whitehorse Trough, and comprises the lower volcanic Povoas formation and the upper sedimentary Aksala formation that contains limestones of the Hancock member (Tempelman-Kluit 1984, 2009; Hart 1997; Colpron et al. 2007). Upper Paleozoic rocks of Quesnellia occur along the eastern flank of the Whitehorse trough where they comprise the volcanic, volcanoclastic, and carbonate rocks of the Boswell assemblage. These rocks are unconformably overlain by Upper Triassic volcanic and volcanoclastic rocks of the Semenof formation, which are also assigned to Quesnellia (Simard 2003). The Big Salmon fault separates the Boswell assemblage from the Upper Devonian and older, mostly metasedimentary Snowcap assemblage of the Yukon-Tanana terrane. The Snowcap assemblage is metamorphosed to amphibolite facies, contains Late Devonian to Early Mississippian granitic intrusions, and is inferred to be a rifted fragment of the western Laurentian margin, upon which Devonian-Permian

magmatic arc rocks of Yukon-Tanana were deposited (Nelson et al. 2006; Piercey and Colpron 2009).

The Lower to Middle Jurassic Laberge Group, which was deposited in the Whitehorse Trough (Fig. 3), is an overlap assemblage that extends across the inferred contacts between the Stikinia, Quesnellia, and Cache Creek terranes (Colpron et al. 2015). The Laberge Group unconformably overlies the Triassic Lewes River Group, and block faulting and erosion of Stikinia prior to deposition of the Laberge Group has been inferred from its contact with different parts of the Lewes River Group (Colpron et al. 2007); the Laberge Group is usually in contact with the Aksala formation, but southeast of Carmacks the Laberge Group immediately overlies the Povoas formation. In the northern Whitehorse trough, the Laberge Group comprises the Tanglefoot formation, which contains shallow marine to fluvial sandstone and conglomerate, together with a small amount of shale containing coal, and the Nordenskiöld facies of volcanoclastic rocks. The Middle Jurassic to Lower Cretaceous Tantalus Formation containing fluvial conglomerate, sandstone and coal disconformably overlies the Laberge Group, and delimits the final stage of deposition in the Whitehorse trough (Long 2005; Colpron et al. 2015). We follow the usage of White et al. (2012), and in this paper restrict the use of Whitehorse trough to the strata of the Laberge Group.

The northern Whitehorse trough is intruded by only a few small granitoid plutons of Middle Jurassic and Early Cretaceous age (Fig. 2; Colpron et al. 2007; Colpron 2011). It is, however, flanked on all sides by voluminous batholiths of Late Triassic to Middle Jurassic age, e.g. the Aishihik and Tatchun batholiths (Fig. 2). These were exhumed during development of the trough, and constitute the main source of sediments for Laberge strata (Colpron et al. 2015). The

Carmacks Group, which represents the erosional remnants of Upper Cretaceous flood basalts that covered most of the northern Whitehorse trough, is found widely across the area.

Faulting and deformation

Near the seismic survey, the Whitehorse trough is bounded to the east by the northeast-dipping Tadru thrust fault that placed the Paleozoic Boswell assemblage over Upper Triassic volcanic rocks of the Semenof formation (Quesnellia). To the west, the Braeburn fault, which is inferred to have dextral strike-slip displacement, marks the western limit of Jurassic strata near Carmacks (Tempelman-Kluit 1984; Colpron et al. 2007, 2016), but projects within the Whitehorse trough to the south (Fig. 2). The Teslin fault is considered to be the northern continuation of the Thibert-Kutcho fault system of northern British Columbia where it represents the contact between Quesnellia and the Cache Creek terrane (Gabrielse et al. 2006). Its northern extension into Yukon is, however, not well constrained. The Cache Creek terrane terminates at $\sim 60^{\circ}45'N$ and the Teslin fault juxtaposes similar Triassic and Jurassic strata further north (Fig. 2). In the northern Whitehorse trough, Colpron et al. (2007) interpreted the Teslin fault at the juxtaposition of the Hancock limestone of the Lewes River Group against clastic strata of the Laberge Group approximately 15 km northeast of Carmacks (Fig. 2). To the south, the Teslin fault may have a history of southwest-vergent thrusting during the Early to Middle Jurassic (Mihalnyuk et al. 2004), but then acted as a dextral strike-slip fault during the middle Cretaceous (Gabrielse et al. 2006) when the fault could also have been active in the northern Whitehorse trough (White et al. 2012). Along the northeast flank of the Whitehorse trough, the Tatchun normal fault drops Laberge Group strata relative to Lewes River Group rocks to the northeast, and further north an inferred splay of the fault juxtaposes Lewes River Group against the Middle Jurassic McGregor pluton (Colpron et al. 2007, 2016).

The present northern Whitehorse trough is dominated by southwest-directed shortening that probably started during the later stages of deposition of the Laberge Group. Northwest-trending folds and southwest-verging thrust faults affect Middle to Upper Jurassic, and possibly Lower Cretaceous strata, with poorly constrained estimates of shortening in the range of 15-20% (White et al. 2012) or 35% (English et al. 2005). Between the Teslin and Braeburn faults, fold and thrust structures are interpreted to exhibit a sigmoidal pattern consistent with dextral transpression controlled by these two faults (Colpron 2011; White et al. 2012), which may have further increased the shortening during the middle Cretaceous when dextral strike-slip motion was most pronounced along the Teslin fault further south (~130 km; Gabrielse et al. 2006). Mid to Late Cretaceous dextral strike-slip displacement is also inferred on the Big Salmon fault (Fig. 2).

2004 Carmacks seismic survey

The seismic survey was designed to use existing roads to cross the northern Whitehorse trough, and continue further east across the adjacent Quesnellia and Yukon-Tanana terranes to determine if Mesozoic sedimentary strata have been overridden. Seismic line 1 is a 117 km east-west oriented line (Fig. 2) along the Robert Campbell Highway that starts near the eastern edge of the Snowcap assemblage in the Yukon-Tanana terrane, crosses the Big Salmon fault, the Tadru fault, and the entire width of the Whitehorse trough before ending 6 km west of the Braeburn fault on the Mount Nansen Road. Line 2 was shot north-south for 53 km along the North Klondike Highway, intersecting line 1 at Carmacks. Line 2 runs along the western edge of the Whitehorse trough, obliquely crossing the major northwest-trending structures, before continuing for 8 km along the Braeburn fault at the line's southern end.

The survey was designed to provide good spatial resolution in the upper 5 km, while still obtaining signal penetration to the base of the crust that could be recovered using an extended correlation technique (White et al. 2006). A source array of three large 22,000 kg vibrators with vibration points (VP) every 60 m used a 24 s linear upsweep of 10-84 Hz. Single 3-component buried sensors were deployed in a 600-channel symmetric split-spread at a 20 m group interval, the crooked line geometry resulting in a maximum offset of approximately 5800 m. With a 24s upsweep and 9 s post-sweep listen time, extended correlation was applied to the uncorrelated recordings to obtain 20 s records, but with a bandwidth that decreased from 10-84 Hz at 9 s to 10-50 Hz at 20 s.

The seismic reflection data were subject to a standard prestack processing sequence by the Geological Survey of Canada, comprising crooked line geometry, refraction statics with a replacement velocity of 4800 m/s, prestack automatic gain control (AGC), velocity analysis, residual statics, normal moveout with 50% stretch mute, and stack (White et al. 2012). Post-stack migrations shown here were obtained by applying a line segment migration algorithm that does not generate any wave-equation migration artefacts to stacks, to which a semblance-weighted coherency filter had been applied (Calvert 2004). The datum for all the seismic displays is 750 m above mean sea level (MSL).

Approximately 1900 shots were recorded along line 1, and 900 shots along line 2. For the tomographic inversion presented here, first arrivals were repicked along both seismic lines. We used a neural network picking algorithm trained to pick the initial peak of the source signature, but the quality of the automated picking was often poor due to the low signal-to-noise ratio,

especially for offsets greater than 3500 m. Therefore, the shot records were manually edited to ensure a consistent set of picks in both the common shot and common offset domains.

First-arrival tomographic inversion

The first arriving waves from a seismic source are turned back to surface receivers by the increase in P-wave velocity with depth. Where there are multiple source and receiver positions along a seismic line, there are many different crossing propagation paths (ray paths) through the subsurface, which in the case of the crooked Carmacks seismic lines are not confined to a single vertical plane. Tomographic inversion of the first arrivals can be used to estimate a distribution of subsurface seismic velocities that will reproduce the observed first arrival travel times. Since the seismic propagation path depends on the velocity model, which is not known, this inversion problem is non-linear, requiring an iterative method that starts from an assumed velocity model that is a reasonable approximation to the actual subsurface velocity variation. The estimated velocity model is non-unique, because it depends on the choice of starting model, and convergence of the iterative method is controlled by constraints on the form of the desired model, e.g. smoothness, and limited by arrival time picking errors. We use a 1-D starting model, in which velocity varies only with depth below ground level, to avoid introducing unnecessary lateral velocity variation into the final model estimate.

Velocity perturbations are best resolved where a large number of seismic propagation paths intersect at a wide range of angles. In parts of the model where propagation paths are largely parallel to one another, resolution can be poor. Resolution is typically evaluated by attempting to recover an arbitrary perturbation, e.g. a checkerboard, superimposed on the final estimated velocity model; a well-recovered perturbation implies good resolution. Zones where there is a

decrease in seismic velocity with depth can be difficult to resolve using first arrival travel time tomography, and volcanic-covered sedimentary basins, where high velocity basalts with P-wave velocities of, for example, 6.0 km/s overlie porous clastic rocks with velocities of 3.0 km/s, fall into this category. Although there is evidence of some sudden first arrival time delays, or “shingling”, related to a velocity decrease with depth under the Carmacks volcanics at the west end of line 1 where the data quality is poor, it appears that the sub-surface is generally suitable for application of the tomographic inversion methodology.

Tomographic inversion method

First arrivals from the two seismic lines were inverted using the 3-D First Arrival Seismic Tomography (FAST) program of Zelt and Barton (1998). In this iterative algorithm, travel times and ray paths are calculated through an initial velocity model, and a regularized update to this model is obtained from the mismatch between the observed and calculated times; for the Carmacks data, a second derivative smoothness regularization constraint was employed. The modelling and update calculation is repeated until the observed travel times are reproduced to within the picking error or a set number of iterations have been carried out.

Definition of model geometry

Typically, first arrival tomography is carried out on a line-by-line basis; however, since Carmacks lines 1 and 2 intersect and to ensure that the velocity models are consistent at the intersection point, a single 3-D velocity model was defined, allowing the two lines to be inverted together, but at the expense of greater computational cost since travel times must be calculated through a largely empty model.

The travel time picks from lines 1 and 2 were exported with UTM coordinates (Zone 8; NAD 83 ellipsoid, central meridian of 135° W) after subtraction of an origin point at (425000, 6786000), and converted to model coordinates in kilometres. The 3-D tomographic model extends from 0 to 100 km in X, -6 to 40 km in Y and -1.0 to 1.5 km in Z, with Z, i.e. depth, increasing downward; depths in the model are referenced to a zero datum of mean sea level (MSL).

Starting model

Seismic tomography is an iterative procedure, in which an update perturbation to a velocity model is calculated from the travel time misfit, and thus requires an initial starting model. To avoid introducing any artefacts or structure into the final velocity model via the starting model, a 1-D velocity variation corresponding to a continuously differentiable exponential function with respect to depth was used for the starting velocity model:

$$V(z) = V_B - (V_B - V_0) \exp[-\alpha(z - z_0)]$$

where V_B is the limiting velocity corresponding to the igneous basement, α is a constant that determines the rate of increase of velocity with depth, and V_0 is the velocity at a user-specified reference depth z_0 . In addition, to ensure the propagation of rays in the sub-surface, and not above ground level, velocities were set to 1200 m/s above the ground level surface in the 3-D velocity model, which was determined from the elevation of the nearest shot point on either of the two seismic lines. This surface is inaccurate away from the seismic lines, but this approximation is sufficient as there is no seismic ray propagation in this part of the model.

The reference depth z_0 was set to -700 m, i.e. 700 m above MSL and just above the highest shot point used for the tomographic inversion. The actual topographic variation in the model is -875 m to -508 m, but points above -700 m correspond to shot points near the western edge of the

model where data quality was poor and no picks could be obtained. V_B was set to 6000 m/s, which is representative of shallow igneous basement. The values for V_0 and α of 3000 m/s and 0.0010 respectively were determined by calculating the root mean square (RMS) misfit between forward modelled travel times and the observed picks for every 100th shot for a range of trial values of V_0 and α .

Iterative inversion

Using the starting model, the 3-D tomographic inversion was run for a total of 1778 picked shots from Carmacks lines 1 and 2. For the forward calculation of the travel times, the velocity model was discretized at 50 m, but for the inversion a 100 m cell size in the X and Y directions and 50 m in the Z direction was used. Second derivative smoothness constraints were specified with equal weighting in horizontal and vertical directions, and the weighting of these constraints relative to the smallest perturbation from the starting model was 99%. The trade-off parameter, which controls the relative fit to the constraints versus the travel time picks, was gradually reduced for subsequent iterations to introduce more short wavelength structure into the inverted model. The RMS misfit between the calculated and observed travel times was reduced from 105 ms with the starting model to 26 ms after four iterations. With subsequent iterations, the RMS misfit increased before decreasing after nine iterations to a final value of 24 ms, but the resulting velocity model contained unrealistic short-wavelength structures and is not presented.

Corrugation model resolution tests

The ability of the tomographic inversion to resolve structural details was evaluated along line 1 using corrugation perturbation tests. The test consists of the addition of a small perturbation to the line's tomographically inverted final velocity model, computation of synthetic travel times by

forward modelling all receivers and sources used in the inversion, and then inversion of these synthetic times using the same parameters as for the field data. The resolution of the recovered model, both laterally and in depth, can then be estimated from how well the perturbation, i.e. the difference between the inversion for the perturbed model and the original final model, is recovered. Three separate corrugation tests using a sinusoidal perturbation that varied only in the Y direction with maximum amplitude of 0.5 km/s and a half-wavelength of 0.8 km, 1.6 km or 3.2 km were carried out. Vertical slices of the inverted 3-D model along the common depth point (CDP) binning line used in the reflection processing are shown in Figure 4. (Each CDP bin extends 10 m along a smoothed line fit to the actual locations at which the survey was acquired). The uniformity with depth of the recovered perturbation provides an indication of how well structures of a given size can be recovered with depth, with the change in polarity of the perturbation indicating the maximum recovery depth. Overall the results of the checkerboard tests indicate that structures with a dimension of 3.2 km are resolved down to a depth below ground level of 0.5-1.0 km, structures with a dimension of 1.6 km are resolved to depths of 0.4-0.8 km, and structures with a dimension of 0.8 km are poorly recovered.

Tomographic velocity models

The seismic P-wave velocities estimated by tomographic inversion provide a general constraint on the variation in subsurface lithology. Although seismic velocities can vary within a particular geological unit, it is often possible to place bounds on expected velocities using measurements on cores, sonic well logs, or correlation with known surface outcrop. Seismic velocity, however, is reduced by higher porosities, which can occur in unconsolidated rocks and in rocks at depths less than 500-1000 m where the confining pressure is insufficient to close microcracks.

Line 1

Line 1 extends from the Yukon-Tanana terrane east of the Big Salmon strike-slip fault, crosses the Whitehorse trough, including rocks of both Quesnellia and Stikinia, and ends west of the Braeburn strike-slip fault in rocks of the Carmacks Group, which overlies Upper Triassic Povoas Formation and Early Jurassic Aishihik batholith to the south (Fig. 2). The final velocity model shows some significant changes in velocity that are reasonably well-constrained to depths of 1 km below ground level.

Close to the surface, there are localised regions up to 400 m thick with velocities of 2.0-3.0 km/s. Between 60 km and 88 km, these low velocity units appear to correlate with outcrops of the Carmacks volcanics (v in Fig. 5a). Although such low velocities are not typically associated with volcanic rocks (values determined from core samples are commonly 4.0-6.5 km/s), first arrival tomography across the Nechako-Chilcotin plateau found velocities of 1.8-3.0 km/s associated with the Miocene-Pliocene Chilcotin volcanics (Hayward and Calvert 2009; Talinga and Calvert 2014). The explanation for such low velocities is unclear, but could be related to the presence of a variety of high porosity volcanic facies, which may in some cases occur above the water table. The regions of near-surface low velocity at 46-50 km and 55-58 km are not correlated with volcanic rocks, and are more likely to be related to less consolidated alluvial deposits in the vicinity of the Little Salmon River (a in Fig. 5a). Further west, near-surface velocities are mostly greater than 3.0 km/s, although relatively low surface velocities at 28 km may correlate with volcanic rocks or alluvial deposits related to the Yukon River (a/v in Fig. 5a). At 40-42 km, velocities of 3.0-4.0 km/s may alternatively be related to fracturing created by the Teslin fault whose location (Fig. 2) has been inferred by Colpron et al. (2007) near 43 km (a/v/f in Fig. 5a).

East of the Big Salmon fault, velocities are ~ 4.0 km/s at the surface, increasing with depth to 5.0 km/s over a few hundred metres. These velocities are broadly consistent with common values for metasiliclastic rocks, which predominate in the Snowcap assemblage that is present here. West of the Big Salmon fault, velocities at depth of 5.5-6.0 km/s likely correspond to igneous basement rocks, and probably represent the Tatchun batholith, which is bounded to the west by the Tatchun fault (Fig. 2). Overlying velocities of 3.5-5.0 km/s might correspond tentatively to volcanic rocks of the Upper Triassic Semenof formation, which outcrops at 51-55 km where velocities are ~ 4.0 km/s.

The Jurassic Laberge Group is the predominant surface lithology from the Tatchun fault to the Hoochekoo fault (Fig. 2). Near-surface velocities are commonly ~ 4.0 km/s increasing with depth to 5.0 km/s over 300-800 m, values which are consistent with relatively low porosity clastic rocks found in other Intermontane basins such as beneath the Nechako-Chilcotin plateau (Talinga and Calvert 2014). The highest surface velocities of 4.2 km/s correlate with an outcrop of limestone of the Upper Triassic Hancock member of the Lewes River Group (c in Fig. 5a), and it is possible that some of the higher velocities at depth are related to the presence of limestone, which may be in faulted contact with Laberge Group clastic rocks.

West of the Hoochekoo fault, the 4.0 km/s isovelocity contour defines a small structural basin up to 800 m deep. The Jurassic-Cretaceous Tantalus Formation is present at the surface over the central part of this feature, and rocks here with velocities of 3.0-4.0 km/s may correspond to this unit, although rocks with velocities at the upper end of this range could also indicate the Laberge Group.

Line 2

Line 2, which lies entirely within Stikinia, extends south from the Tatchun fault, obliquely crosses the northern tip of the Whitehorse trough, and runs for 8 km along the Braeburn fault at the line's southern end (Fig. 2).

Near-surface zones of relatively low velocity, 2.0-3.0 km/s, also exist along line 2, but only extend at most to depths of 200 m below ground level. At 0-2 km, 13-15 km, and 30-34 km, these low velocities occur where the seismic line approaches the Yukon River, and the velocities probably indicate the presence of alluvial deposits (a in Fig. 5b). Velocities of 2.0-3.0 km/s at 20-27 km correlate with Carmacks volcanic rocks (v in Fig. 5b), as also observed on line 1. Where the southern end of the seismic line runs along the Braeburn fault, Carmacks volcanics are present immediately east of the fault, but the line is also adjacent to the meandering Nordenskiöld River where alluvial deposits have probably accumulated (Fig. 2). Elevated fracture porosity in the fault zone could additionally reduce seismic velocities, and all three factors could contribute to the low velocities here, but they do not extend below 200 m.

North of 20 km, beneath the Laberge Group, the 4.0 km/s isovelocity contour occurs at ~ 200 m below ground level, and velocities increase with depth to 5.0-5.3 km/s near the base of the constrained velocity model, consistent with values observed on line 1; however, the 5.0 km/s contour exhibits significant topography. Potentially the relatively high velocities of 5.0 km/s at ~400 m depth could indicate block faulting and/or the presence of carbonate rocks (c in Fig. 5b); 3 km southeast of line 2, for example, limestone rocks of the Hancock member occur at the surface immediately east of the Teslin fault (Fig. 2; Colpron et al. 2007).

From 20-36 km, there is a 600-1000 m thick unit where velocities increase from 3.0-4.5 km/s. The Tantalus Formation could account for much of this thickness, because these rocks are present at the surface from 27-36 km, and probably extend beneath Carmacks volcanic rocks as far north as 24 km, and perhaps as far as 20 km; a similar thickness of rocks with these velocities was also observed on line 1. The depth of the transition to the underlying Laberge Group probably occurs at a depth below ground level of 700-1000 m where velocities increase to >4.5 km/s, values which occur at a much shallower depth below the outcropping Laberge Group to the north. At the southern end of line 2, relatively high velocities of ~5.5 km/s are inferred at depths of ~500 m below ground level, and likely indicate the presence of relatively shallow, possibly granitic rocks.

Reflection images

To correlate the P-wave velocity models with the coincident reflection images, the models for lines 1 and 2 were shifted to a horizontal datum of 750 m above MSL using the same replacement velocity of 4800 m/s that was employed in the processing of the reflection data, converted from depth to two-way travel time, and superimposed on migrated seismic reflection sections. The seismic data quality varied along the two seismic lines, most likely due to surface conditions, and in areas of reduced data quality lower crustal reflections were not apparent in vertical bands on the unmigrated sections; upper crustal reflections did not seem to be significantly affected. The zones of lower data quality are not as apparent in the migrated images, but these zones are indicated on the displays (Figs. 6 and 7). Overall the application of coherency filtering reveals more arrivals in the upper crust than were apparent in the original seismic interpretations (White et al. 2012).

Although some of these events may be coherent noise or out-of-plane arrivals, it is possible to identify discontinuities and changes in the apparent dip of reflection packages, from which the position of faults are inferred; true reflector dips cannot be readily estimated from a single 2-D line. Changes in orientation of the seismic lines are annotated at the top of Figures 6 and 7. The largest change in direction of 90° occurs at CDP 6900 along line 1, but no clear changes in apparent reflection dip can be associated with this change in direction due to the lower data quality here and an absence of laterally continuous reflections. Specific features are interpreted in more detail below, but there is inevitably some ambiguity, particularly where crossing arrivals are present; however, the distribution of inferred faults delineates major crustal blocks, and provides an indication of the overall structural style of the Whitehorse trough and Intermontane terranes in Yukon.

Line 1

Upper crust

From CDP 1-1000, reflections at 0-2 s define a broad antiform, into which overlying reflections flatten, but immediately west of CDP 1200 reflections dip more steeply to the east, suggesting that two distinct crustal blocks within the Snowcap assemblage may be separated by a fault that projects to the surface near CDP 1500 (F1 in Fig. 6). The east-dipping package of reflections is truncated at CDP 2300-2700 by high amplitude west-dipping reflections, which are interpreted to overlie a fault F2 that projects to the surface at CDP 2250, close to the mapped location of the Big Salmon fault; below 2 s, F2 can be extended to depth as the truncation of east-dipping reflectors in its hanging wall. The west-dipping reflections project upward into the near-surface low-velocity zone associated with the Carmacks volcanic rocks between CDP 2300 and CDP

2700, raising the possibility that these dipping reflectors are Upper Cretaceous sills associated with emplacement of the volcanics. The west-dipping reflections appear to be separated by another fault from an upper crustal block between CDP 2800 and CDP 4500 that contains subhorizontal to east-dipping reflections, which project into Carmacks volcanic rocks at CDP 3600-4200 and might also be interpreted as sills. Sill emplacement may be controlled by faults and/or stratigraphy, but sills that cut across existing structures introduce further uncertainty into any interpretation.

No seismic reflection can be clearly associated with the surface location of the Tadru fault, although a fault, F3, which projects to the surface at CDP 4500, is tentatively inferred from a lateral change in apparent reflector dip. Between CDP 4500 and CDP 5600, most reflections at 1-4 s dip to the east, but reflections west of CDP 5600 dip to the west, suggesting that a sub-vertical fault F4 may be present beneath the mapped location of the Tatchun fault. Although the seismic line changes its orientation near this point, it is only by at most 45° , and unlikely to cause this change from easterly to westerly apparent dip. The interpreted sub-vertical fault could flatten into a west-dipping fault at 2-4 s, or perhaps into the deepest east-dipping reflection at 2-4 s below the surface location of the Quesnellia terrane (Fig. 6). Alternatively, the Tatchun fault may correspond to fault F5 interpreted above a small package of moderately dipping reflections just west of F4, and the relatively low velocities between CDP 4600 and CDP 6200 might represent a small structural basin in the hanging wall, consistent with the normal fault mapped at the surface (Colpron et al. 2007).

The Teslin fault, which is mapped as a northeast-dipping thrust (Colpron et al. 2007) and/or dextral strike-slip fault (Colpron 2011; Colpron et al. 2016), is shown at CDP 6400, although

there is some uncertainty in its exact position. If the fault corresponds to the east-dipping package of reflections, F6, that truncates underlying reflections, and reaches the surface at CDP 6800, then the fault is shallow, being truncated in turn at 1 s. As discussed later, the Teslin fault is a major upper crustal fault within the northern Cordillera, and may in fact correspond to the sub-vertical Tatchun fault as currently mapped, an interpretation more in line with the map of Tempelman-Kluit (1984). Further west, there are no reflections or discontinuities that can be associated with the Hoochekoo fault due to the presence of several cross-cutting arrivals that may be coherent scattered noise. More generally, between the Tatchun and Braeburn faults, reflection packages above 2 s exhibit a variety of dips, and can potentially be separated by faults, which divide the uppermost crust into a number of distinct blocks that are typically only a few km across (Fig. 6). Below 2 s, these inferred faults are either listric, or are truncated by listric faults, and they define a basin-like geometry. One of these listric faults, F7, projects towards the mapped location of the Braeburn fault, but the reflections cannot be continued to the surface due to poor data quality at the western end of line 1, which also meant it was impossible to pick first arrivals to use for the tomographic inversion. In general, the seismic reflection section provides no constraint on the sense of motion of the interpreted faults, although the tomographic velocities may provide an indication, as noted above in the case of the Tatchun fault.

Middle and lower crust

Lower crustal reflections at 7-11 s are subhorizontal to shallow dipping; they exhibit lower amplitudes and less lateral continuity in zones of reduced data quality, but also at CDP 7500 (S1 in Fig. 6), which is not clearly associated with a data quality problem. Between CDP 1 and CDP 500, middle crustal reflections are mostly subhorizontal, and are truncated between CDP 3500 and CDP 5000 by the downward continuation of fault F2. Near CDP 10000, a 0.5-1.0 s thick

package of reflections, F8, dips east, but flattens out at 7 s near CDP 8000 where F2 appears to merge into its upper surface. Together, F2 and F8 define a broad synform that underlies the more complex upper crustal geometry of the Whitehorse trough. S1 occurs at the base of the synform, and we tentatively suggest that the reduction in lower crustal reflectivity here indicates a subvertical strike-slip shear zone, which could represent the stem of a strike-slip flower structure that includes F2 and F8; however, given the data quality issues, interpretation of S1 as a fault is uncertain.

Line 2

Upper crust

At the northern end of line 2, the crooked geometry results in the seismic line being oriented almost parallel to the main northwest-striking structures. The Tatchun fault correlates with reflections, J1, that exhibit an apparent south-southeast dip from CDP 1 to CDP 300 (Fig. 7). The Teslin fault appears to correlate with approximately northeast-dipping reflection, J2, which projects to the surface at CDP 1100. These reflectors flatten out at 1 s, indicating that the Teslin fault may not extend below 4 km depth, as also suggested by the interpretation of line 1. North-northeast dipping reflections, J3, project to the surface close to the Hoochekoo fault, but also appear to flatten or be truncated by subhorizontal reflections at approximately 2 s. These reflectors may represent the fault, or perhaps sills associated with the Carmacks volcanics.

Between CDP 300 and 2800, deeper reflections are mostly shallowly dipping, consistent with gentle folding, but may also define a synform with its centre at CDP 1700. At the south end of line 2, the reflectivity is mostly subhorizontal, consistent with the seismic line running along

strike, and also along the Braeburn fault. But near CDP 3400 a package of south dipping reflections are truncated by reflectors with an apparent dip to the north, suggesting that a fault, J4, may be present near the southern limit of the exposed Tantalus Formation.

Middle and lower crust

North-dipping upper crustal reflections do not continue below 4-5 s, and they may be truncated by fault J5, which correlates with some weak, intermittent north-dipping reflections. Well-defined subhorizontal reflections are present from 7 s to 11 s at CDP 1-1000 and CDP 4200-4500, but the lower crustal reflectivity appears weaker in other areas, in large part due to the noted poor data quality. There is no strong constraint of the existence or otherwise of the possible continuation of fault S1 inferred on line 1, but, if it exists, this subvertical shear zone could be present anywhere between CDP 1800 and CDP 3800 (Fig. 7).

3-D Gravity Inversion

To provide further constraint on lithological variation beneath the tomographic velocity model, the Bouguer gravity anomaly was inverted in 3-D for a crustal density model. High-resolution gravity data (~2 km station spacing) acquired by helicopter airborne surveys (Yukon Geological Survey 2011, 2012) in the region of Carmacks were merged with regional (~10 km station spacing) ground-based gravity data (Fig. 8a), extracted from the Geophysical Data Repository (Geological Survey of Canada 2014). Gravity data were reduced to the Bouguer anomaly using a standard crustal density (2.67 g/cm^3), and a constrained inversion was performed using the GRAV3D software package (Li and Oldenburg 1998). The inverse problem was solved by minimizing the data misfit between the observed and predicted anomaly subject to model

constraints; a trade-off parameter was used to balance the data fit and model smoothness (Li and Oldenburg 1998).

The model for the inversion was defined on a mesh with a core cell size of 5 x 5 x 1 km with the surface topography defined on a 5 km topographic grid. The input Bouguer gravity data was gridded at a spacing of 5 km using a continuous curvature algorithm in Oasis Montaj (Geosoft Inc.), and upward continued by 500 m to minimize shallow-sourced perturbations and noise. These preprocessing steps and the inversion parameters were chosen to closely reproduce the observed anomaly, whilst balancing model resolution, smoothness, artefacts, and run-time. In order to constrain the inversion to place most density variations within the crust, the inversion was constrained by a simple two-layer reference model separated at a Moho depth of 35 km (for example Cook et al. 2004; Fernández-Viejo et al. 2005; see also Figs. 6 and 7) with density contrasts of 0.1 g/cm³ above and 0.0001 g/cm³ below. The calculated gravity anomaly closely matches (RMS = 1.01 mGal) the observed anomalies (Fig. 8). Since it was necessary to utilise smoothness constraints to address the non-uniqueness inherent in gravity inversion, it is possible that some of the sub-vertical contrasts in the lower crust are due to the smearing of shallower features into the deeper parts of the model.

The density variation relative to the reference model along seismic line 1 has been extracted from the 3-D density model, and is superimposed on the seismic interpretation in Figure 6b. Fault F1 corresponds to a reduction in density (~0.1 to 0.04 g/cm³) towards the western edge of the Yukon-Tanana terrane. In the Quesnellia terrane, lower densities (~0.05 to 0.09 g/cm³) occur between faults F2 and F3, and correlate with the Tatchun batholith, whereas high densities

(~ 0.13 to 0.17 g/cm^3) are present between CDP 4200 and CDP 5800, indicating more mafic igneous rocks of the Semenof formation that extend to $>10 \text{ km}$ depth. Relatively high-density rocks (~ 0.11 to 0.12 g/cm^3) are also inferred within Stikinia west of CDP 9700; these probably correspond to basalt of the Povoas Formation (Fig. 2). Above 4-5 km depth between CDP 6500 and CDP 9500, lower modeled densities of ~ 0.09 to 0.1 g/cm^3 probably indicate the extent of the clastic rocks within the Whitehorse trough where a number of the more steeply dipping faults are interpreted. The base of this low density region corresponds approximately to the shallower of the low angle faults, for example F7 and the westward continuation of F5, which may therefore indicate the position of the top of the igneous basement. A similar transition to lower densities within the basin is also found between CDP 1000 and 3000 on line 2 (Fig. 7b).

The seismic interpretation of F2 and F9 appears to be broadly consistent with the inverted density model (Fig. 6b). Within the middle crust, the downward projection of the Big Salmon fault, F2, correlates approximately with the decrease in density with depth at CDP 4000-6000 below Quesnellia. Fault F9, which projects into the Tummel fault along the western edge of the Slide Mountain terrane, flattens westward at $\sim 12 \text{ km}$ depth where there is a density high along the fault trace and a downward reduction in the density gradient. S1 is tentatively interpreted on line 1 (Fig. 6b) where there is only a subtle change in the horizontal density gradient, with a westward reduction of density in the lower crust; however, on line 2 (Fig. 7b) S1 correlates with a localized reduction in density in the lower crust.

Discussion

The term Whitehorse trough originally referred to the Mesozoic depocentre in which the sedimentary strata of the Triassic Lewes River Group and the Jurassic Laberge Group were laid

down (e.g., Wheeler 1961; Hart 1997). More recently, however, with the identification of a regional unconformity between these two units (Lowey 2008), White et al. (2012) and Colpron et al. (2015) associated the Whitehorse trough only with the Laberge Group (Fig. 3), which is an overlap assemblage deposited on both Stikinia and Quesnellia, as well as the Cache Creek terrane further south, and we have followed this usage. Because the seismic and gravity data cannot discriminate between sedimentary strata of the Lewes River and Laberge groups, and since these units are overlain by younger sedimentary rocks of the Tantalus Formation, we discuss below the distribution of the overall package of Triassic to Lower Cretaceous sedimentary strata found within the present-day structural basin delineated by the seismic reflection survey, which overlies an igneous (and metamorphic?) basement (Povoas basalt and older rocks).

The western edge of the Triassic to Early Cretaceous sedimentary succession is defined in both seismic velocity models by a northeastward increase in depth of the 5.0 km/s isovelocity contour near the Braeburn fault (Fig. 5). Along seismic line 1, the eastern flank of the basin is marked by the gradual shoaling of this contour from the Tatchun fault to the Tadru fault where the Laberge Group is relatively thin and the Tatchun batholith is close to the surface. To the north and south of the seismic line, however, Triassic to Early Cretaceous sedimentary rocks only occur west of the Tatchun fault, which appears to delimit the deeper part of the basin in these areas. In the reprocessed reflection section of line 1 above 2 s, there is a sudden lateral change in reflector dip across CDP 5600, indicating that the Tatchun fault is probably subvertical, supporting the interpretation of this fault as the most consistent eastern boundary to the basin along strike. Between the Braeburn and Tatchun faults, rocks with velocities less than 5.5 km/s, consistent with sedimentary lithology, extend to at least 800-1200 m depth (Fig. 5). In this same region, we

have traced several near-surface reflections, which are likely to be from sedimentary strata, downward to ~ 2 s, corresponding to 5 km depth (Fig. 6a). Several distinct packages of upper crustal reflections are bounded by interpreted faults that either flatten or are truncated by listric faults at 2.5 s, which probably marks the maximum depth of sedimentary rocks, and is generally consistent with the depth estimate of 6-8 km from the gravity inversion (Fig. 6b). Since these strata are cut by several faults, it is possible, however, that in some parts of the basin crystalline basement may locally occur at shallower depths.

Most of the sedimentary rocks in the structural basin have seismic velocities >4.0 km/s, and likely correspond to the Laberge Group. Although clastic rocks of the Lewes River Group are probably indistinguishable from the Laberge Group on the basis of velocity, limestones of the Hancock member may correspond to velocities >5.0 km/s, particularly in the upper 1000 m, where the velocity contrasts are most pronounced due to open microcracks and fracture porosity. A median measured density for the Laberge Group of 2.55 g/cm^3 (Tempelman-Kluit and Currie 1978) corresponds by Gardner's relation to a P-wave velocity of 4.60 km/s, which is broadly consistent with our tomographic velocity estimates. The younger Tantalus Formation can generally be distinguished from the underlying Laberge Group by its lower velocity of 3.0-4.0 km/s, which probably indicates greater average porosity. Somewhat surprisingly, in several locations volcanic rocks of the Carmacks Group appear to exhibit anomalously low velocities of 2.0-3.0 km/s, and this observation is also likely to be due to high porosity.

The seismic reflection data reveal that the sedimentary basin is part of a larger, 18-20 km deep synform that extends to the contacts of Stikinia and Quesnellia with the Yukon-Tanana terrane in the west and east respectively. This synform, named here the Northern Intermontane synform, is

underlain by listric fault F2, which is the downward and westward continuation of the Big Salmon fault, and F8, which projects upward towards the western contact of Stikinia with the Yukon-Tanana terrane (Fig. 6). We thus infer that both Stikinia and Quesnellia, which comprise the bulk of the synform, overlie the Yukon-Tanana terrane, and are separated from it by faults rather than a stratigraphic boundary (Fig. 9). In northern British Columbia, the Thibert fault (equivalent of the Teslin fault in Yukon) marks the contact between Quesnellia and the Cache Creek terrane. To the north, beyond the northernmost exposure of the Cache Creek terrane, the Teslin fault is inferred to mark the boundary between Stikinia and Quesnellia (Fig. 2). Near Carmacks both seismic reflection lines indicate that the Teslin fault as mapped by Colpron et al. (2007) does not extend below 1 s, approximately 3 km depth. We suggest that here the Tatchun fault is the boundary between Stikinia and Quesnellia, because it continues to greater depth and marks a distinct lateral change in reflector dips near CDP 5600; the upper crustal block between the Tatchun and Big Salmon faults is distinguished from upper crust to the west by a consistent set of east-dipping reflections. If the Tatchun fault dips to the east, then Quesnellia is less than 14 km thick, and does not underlie Stikinia (Fig. 9). Alternatively, if the fault flattens to the west, Quesnellia rocks may extend as deep as 20 km beneath the axis of the Whitehorse trough. Given its greater depth extent, we suggest that the Tatchun fault is more likely to be the continuation of the Teslin fault mapped in southern Yukon. Due to the similarity in their lithology, Stikinia and Quesnellia may in fact be two fault-bounded blocks derived from a single terrane that have been juxtaposed along the Tatchun/Teslin fault system.

Although the thickness of the Yukon-Tanana terrane is unclear, the near-surface block in the footwall of the Big Salmon fault appears to extend to at least 5 s (Fig. 6). The reflections interpreted as fault F9, which project towards the Tummel fault and a small sliver of the Slide

Mountain terrane 15 km beyond the east end of line 1 (Colpron et al. 2005), mark the base of the Yukon-Tanana terrane, and also flatten at ~5 s. The seismic lines are insufficiently long to provide any clear understanding of the origin of the underlying lower crust at 7-11 s (Figs. 6 and 9). In northern British Columbia, seismic data have been interpreted to indicate that the Insular terranes were thrust beneath the Intermontane terranes during the Cretaceous (Cook et al. 2005), and this may also be the case under the Whitehorse trough, which would explain the apparent thinning of the Yukon-Tanana terrane beneath the synform. However, the subhorizontal lower crustal reflectivity, which is well imaged where the data quality is high, is similar to that observed beneath the Nechako-Chilcotin plateau in British Columbia, and may indicate reworking of the lower crust by extension and basaltic intrusion (Calvert and Talinga 2014), perhaps associated with the Cretaceous volcanism that produced the Carmacks flood basalts. Due to the likely reworking and possible displacement, of the lower crust under the Whitehorse trough, we are unable to determine whether the present crustal structure is consistent, or otherwise, with oroclinal bending during the Jurassic accretion of the Intermontane terranes. Beneath the Cassiar terrane, the lower crust is probably of North American affinity (Fig. 9), but the western part of this lower crustal block may have resulted from northeastward displacement of accreted terranes during the Cretaceous.

Given the eastward change across the Big Salmon fault from sub-greenschist to amphibolite metamorphic facies, we suggest normal motion dropped the Quesnellia and Stikinia terranes into the crustal-scale Northern Intermontane synform during orogenic collapse in the middle to Late Cretaceous (Staples et al. 2016), resulting in the terranes' preservation. With dextral transpression along much of the Cordillera during and after this time, it is likely that many interpreted faults exhibit a significant component of strike-slip motion, which is consistent with

our interpretation of multiple upper crustal blocks bounded by relatively small faults in the Whitehorse trough. Faults within the sedimentary basin are bounded below by two listric faults, one of which could be the Tatchun/Teslin fault (F4 on Fig. 6). These two faults terminate at 6-7 s depth, close to inferred shear zone S1, which may be the base of a crustal-scale flower structure, indicating a zone of distributed strike-slip deformation within the lower crust that probably erased any pre-existing seismic reflectivity. Although thrust faults in the western part of the Whitehorse trough could indicate a positive flower structure, this inference is likely an oversimplification of a complex history of strike-slip faulting that probably dates back to the Middle Jurassic (Nelson et al. 2013). In the absence of other evidence for strike-slip deformation near the base of the crust, the interpretation of S1 is speculative, and it is possible that strike-slip movement is restricted to the upper and middle crust, which may have been mechanically decoupled from the lower crust.

Conclusions

The Whitehorse trough is an asymmetric northwest-striking Mesozoic sedimentary basin, which may be as deep as 5-8 km on its western flank based on 3-D inversion of Bouguer gravity data. Tomographic inversion of first arrivals recorded in a seismic reflection survey across the basin indicates that Upper Jurassic to Lower Cretaceous sedimentary rocks of the Tantalus Formation have seismic velocities <4.0 km/s and can generally be distinguished from the Jurassic Laberge and Triassic Lewes River groups, which are inferred at depths up to 1200 m and commonly have velocities of 4.0-5.2 km/s. Low velocities of 2.0-3.0 km/s may indicate volcanic rocks of the Carmacks Group, alluvial deposits, or in some cases fault zones.

Deep seismic reflection data indicate that the sedimentary basin is underlain by Stikinia in the west and Quesnellia in the east, and that these two terranes form the broader crustal-scale Northern Intermontane synform that is in turn underlain by the Yukon-Tanana terrane, based on the continuation to depth of the terrane bounding faults. A localized absence of reflectivity in the lower crust beneath the axis of the basin may be due to a subvertical strike-slip shear zone that is linked to the terrane bounding faults, controlling strike-slip faulting in the upper crust as part of a broad flower structure.

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Figure Captions

Fig. 1. Intermontane sedimentary basins (black dashed lines) superimposed on major tectonostratigraphic terranes of the Canadian Cordillera: YT – Yukon-Tanana, ST – Stikinia, QN – Quensellia, CC – Cache Creek, SM – Slide Mountain, CA – Cassiar, KS – Kluane, Windy, Coast, AX – Alexander, WR – Wrangellia, OK – Okanagan, AA – Arctic Alaska. NAb – North America deep water sedimentary rocks, NAp – North America platform sedimentary rocks, NAc – North America craton (after Colpron and Nelson 2011). The Carmacks study area shown in Figure 2 is indicated by the solid grey rectangle, and the region of inverted gravity data in Figure 8 by the dashed grey rectangle. Major faults are shown by solid black lines.

Fig. 2. Carmacks seismic lines, with annotated common depth points (CDP) superimposed on geological map of study area (after Colpron et al. 2016). Major units discussed in the text include: 1) Lewes River Group (Stikinia): uTP – Povoas formation, uTA – Aksala formation (h indicates limestone of Hancock member); 2) the Boswell assemblage (uCB) and Semenov formation (uTJS) of Quesnellia; 3) the Laberge Group (JL), including volcanoclastic rocks of the Nordenskiöld facies (IJN) and overlying Tantalus Formation (JKT); 4) uKC – Carmacks Group; and 5) PDS – Snowcap assemblage, MSR – Paleozoic granitoid rocks of the Yukon-Tanana terrane. For other map units consult legend of Colpron et al. (2016).

Fig. 3. Stratigraphic chart for Stikinia (modified from Hart 1997, Lowey 2004, and White et al. 2012). The Semenov and Shonectaw formations of Quesnellia are similar in terms of lithology and age to the Lewes River Group of Stikinia. The Whitehorse trough is considered to correspond to strata of the Laberge Group.

Fig. 4. Corrugation test for line 1 using sinusoidal perturbations that vary only in Y-direction (east-west). Recovered sinusoidal perturbations of half-width a) 0.8 km, b) 1.6 km, and c) 3.2 km. The depth to which the vertical band extends provides an indication of the depth to which the anomalies can be recovered.

Fig. 5. Final inverted velocity models with interpretation and surface geology: a) Line 1, b) Line 2. Note that the location of the Teslin fault shown here corresponds to Figure 2 (Colpron et al. 2016), but the Teslin fault is later reinterpreted in this paper to correspond to the Tatchun fault on the basis of the seismic reflection data. See Figure 2 for geological legend. Vertical exaggeration is 5:1.

Fig. 6. Line 1. a) Interpretation of migrated seismic reflection section with superimposed tomographic velocity model; b) Gravity inversion showing difference from reference model superimposed on interpretation of seismic data. Dashed black lines – interpreted faults with confidence indicated by line length (see text for discussion of F1 – F9); dashed blue lines – inferred major basin bounding faults; dashed red line – possible lower crustal strike-slip shear zone S1; dashed grey line – reflection Moho interface. YTT – Yukon-Tanana terrane; SMT – Slide Mountain terrane. See Figure 2 for geological legend. Changes in orientation of the seismic line are annotated across the top of the figure. Vertical exaggeration is 1:1 for a velocity of 6.5 km/s.

Fig. 7. Line 2. a) Interpretation of seismic reflection section with superimposed tomographic velocity model; b) Gravity inversion showing difference from reference model superimposed on interpretation of seismic data. Dashed black lines – interpreted faults with confidence indicated

by line length (see text for discussion of J1-J5); dashed blue lines – inferred major basin bounding faults; dashed red line – possible lower crustal strike-slip shear zone; dashed grey line – reflection Moho interface. See Figure 2 for geological legend. Changes in orientation of the seismic line are annotated across the top of the figure. Vertical exaggeration is 1:1 for a velocity of 6.5 km/s.

Fig. 8. (a) Bouguer gravity anomaly gridded at 5 km after upward continuation by 500 m. (b) Bouguer gravity anomaly calculated from inverted 3-D model of density contrast. The RMS misfit between the observed and predicted gravity anomalies is 1.01 mGal. Blue box in A shows high-resolution gravity data surveys (Yukon Geological Survey 2011, 2012). Heavy black lines show seismic reflection lines. White dashed lines show selected faults: BF – Braeburn; BSF – Big Salmon; HF – Hoochekoo; TF – Teslin; TNF – Tachun, TUF – Tadru.

Fig. 9. Inferred geometry of the Northern Intermontane synform containing the Stikinia and Quesnellia terranes above the underlying Yukon-Tanana terrane. Within the synform, the sedimentary rocks mostly comprise the Laberge Group, equivalent to the Whitehorse trough, with Stikinia and Quesnellia in contact along the Tatchun/Teslin fault. The Yukon-Tanana terrane was thrust over the Cassiar terrane, trapping a sliver of the oceanic Slide Mountain terrane in a 3-5 km wide fault zone. The origin of the lower crust is poorly resolved, because it was probably reworked by extension and intrusion that produced a ~10 km thick subhorizontal reflection fabric that was also likely affected by localized strike-slip deformation. See Figure 2 for geological legend.

Fig 1: Map of Cordillera

155x233mm (600 x 600 DPI)

184x208mm (300 x 300 DPI)

82x178mm (300 x 300 DPI)

233x131mm (300 x 300 DPI)

249x123mm (300 x 300 DPI)

306x529mm (300 x 300 DPI)

194x101mm (300 x 300 DPI)

792x269mm (300 x 300 DPI)