1 **Surface-Wave Images of Western Canada: Lithospheric Variations**

2 **Across the Cordillera/Craton Boundary**

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16 **Abstract**

17   Two-station surface-wave analysis was used to measure Rayleigh-wave phase velocities

18 between 105 station pairs in western Canada, straddling the boundary between the tectonically

19 active Cordillera and the adjacent stable craton. Major variations in phase velocity are seen

20 across the boundary at periods from 15 to 200 s, periods primarily sensitive to upper-mantle

21 structure. Tomographic inversion of these phase velocities was used to generate phase-velocity

22 maps at these periods, indicating a sharp contrast between low-velocity Cordilleran upper mantle

23 and high-velocity cratonic lithosphere. Depth inversion along selected transects indicates that the

24 Cordillera/craton upper-mantle contact varies in dip along the deformation front, with cratonic

25 lithosphere of the Taltson province overthrusting Cordilleran asthenosphere in the northern

26 Cordillera, and Cordilleran asthenosphere overthrusting Wopmay lithosphere further south.

27 Localized high-velocity features at sub-lithospheric depths beneath the Cordillera are interpreted

28 as Farallon slab fragments, with the gap between these features indicating a slab window. A

29 high-velocity feature in the lower lithosphere of the Slave province may be related to Proterozic

30 or Archean subduction.

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32 **Keywords**: Seismology, surface waves, tomography, Cordillera, craton

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1. Introduction

36   The lithosphere beneath stable continental regions typically exhibits elevated seismic

37   velocities to depths of 200 km or more, which has been attributed to a combination of low

38   temperature and depleted composition (e.g. King 2005, Artemieva 2009, Aulbach 2012). By

39   contrast, young orogenic belts are underlain by low-velocity upper mantle, attributed to a

40   combination of high temperature, elevated volatile content, and a component of partial melt (see

41   e.g. Hyndman et al. 2009). In North America, these two lithospheric regimes exist in close

42   juxtaposition where the Phanerozoic Cordilleran Orogen abuts on Precambrian ancestral North

43   America (Figure 1). Previous tomographic studies of North America (van der Lee and Nolet

44   1997a, Frederiksen et al. 2001, Bedle and Van der Lee 2009, Schaeffer and Lebedev 2014)

45   image the Cordillera-craton transition as a major lithospheric velocity contrast, but questions

46   remain regarding the sharpness of the transition, its orientation in three dimensions, and its

47   relationship to crustal boundaries. We present a new study of the upper mantle beneath western

48   Canada, based on two-station surface-wave analysis, that brings new constraints to bear on these

49   questions.

50   East of the Cordillera, Western Canada is a Precambrian craton assemblage accreted in

51   the Proterozoic from Archean and younger blocks (Whitmeyer and Karlstrom 2007). The Slave,

52   Superior, Rae, Hearne, and Sask cratons are all of Archean age. These cratons assembled during

53   the Proterozoic via the Taltson, Wopmay, and Trans-Hudson orogens. Their western boundary

54   was a passive rifted margin in the Late Proterozoic (Gabrielse and Yorath 1991), onto which a

55   series of terranes were accreted, beginning with the Intermontane Superterrane in the Jurassic

56   (Clowes et al. 2005). This collision produced a fold-and-thrust belt extending to the current

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57 Cordilleran deformation front. A later collision of the Insular Superterrane in the Cretaceous

58 extended the Cordillera westward to its present extent.

2. Data and analysis

60 The Geological Survey of Canada (GSC) operates the Canadian National Seismograph

61 Network (CNSN), which has been recording digital broadband data since the early 1990s (North

62 and Basham 1993). Further instrumentation was installed as part of the Portable Observatories

63 for Lithospheric Analysis and Research Investigating Seismicity (POLARIS) project (Eaton et al.

64 2005), the data from which is also archived by the GSC. We collected data from CNSN and

65 POLARIS stations in western and central Canada, corresponding to earthquakes of magnitude

66 greater than 6, at any distance from the receivers, whose propagation paths lay within 5° of the

67 paths between selected station pairs (Figure 2). These pairs were selected to give consistent

68 coverage of western Canada, and do not include all possible station pairs, particularly in western

69 British Columbia where stations are more closely spaced than is appropriate for the intended

70 resolution of this study. The downloaded data were inspected for Rayleigh-wave signal quality

71 before use. The station responses were removed by conversion to displacement before further

72 analysis; all stations used are broad-band, with all but two stations (ILKN and VGZ) having

73 significant velocity response (at least 10% of the peak value) to periods of 200 s or longer. A plot

74 of instrument responses is provided as supplemental material.

75 Two-station surface-wave analysis is a widely-used technique with a long history

76 (beginning with Sato 1955) in which the change in waveform between two receiving stations is

77 measured. Under the assumption that the waveform change is determined entirely by the

78 structure along the great-circle path connecting the station pair, the waveform change may be

79 used to isolate structure between the receiving instruments, even in aseismic regions. Thus, two-

80   station analysis is appropriate for Canada, which has limited seismic activity away from the west

81   coast; previous studies have successfully applied this approach to the Canadian Arctic

82   (Darbyshire 2005), Hudson Bay (Darbyshire and Eaton 2010), the Superior Province (Darbyshire

83   et al. 2007), the southern Cordillera (Bao et al. 2014), and the northern Cordillera (McLellan et

84   al. 2018).

85   Two-station analysis is generally applied at large distances from the earthquake source.

86   Under these conditions, the different surface-wave modes are well separated and the fundamental

87   Rayleigh mode dominates vertical-component recordings. A single mode's propagation may be

88   described by a dispersion curve showing the relationship between frequency and velocity; the

89   velocity considered may be either group velocity (velocity of energy transport, measured by

90   determining the travel time of the signal envelope peak) or phase velocity (velocity of the phase

91   of a single frequency, measured by determining the phase shift over a given distance). As the

92   group velocity dispersion curve may be determined from the phase velocity, but not vice versa,

93   the phase velocity contains more information (see e.g. Stein and Wysession 2003).

94   We measure phase-velocity dispersion curves using the cross-correlation technique of

95   Meier et al. (2004); an example is given in Figure 3. Two vertical-component recordings of the

96   same event, at stations that lie within 5° of a common great-circle path to the source, are cross-

97   correlated. The phase of the cross-correlation corresponds to the phase difference between the

98   two traces, with an ambiguity of $2\pi n$ radians, where n is an integer. For a cross-correlation phase

99   of Φ measured between two stations separated by a distance x at angular frequency ω, the phase

100   velocity will be

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$$c(\omega) = \frac{x\omega}{\phi + 2\pi n}$$

102   and therefore multiple velocities are compatible with the measurement, the possible solutions

103   being more closely spaced at higher frequencies. In practical use, the distance x is taken to be the

104   difference between the source-receiver distances for each station, to avoid biased measurements

105   for events not perfectly aligned with the two-station path. At sufficiently low frequency, there is

106   likely to be only one phase-velocity value that falls within a realistic range; thus, the approach of

107   Meier et al. (2004) is to begin at the longest period with significant coherent energy, and

108   continue the dispersion curve to higher frequencies while assuming that the curve is smooth (as

109   large jumps in phase velocity are unphysical). Multiple events were analysed for each station

110   pair, incorporating paths travelling in both possible directions, and the resulting set of dispersion

111   curves was averaged to obtain a single high-quality curve for each station pair. Two stages of

112   quality control were performed: on individual paths, events were included in the average only in

113   frequency bands that showed consistency with other events, and on the averaged curves, paths

114   were edited to exclude frequency bands with strong deviations (more than 5%) from the average

115   of all paths.

116   **3. Results**

117   **3.1 Dispersion curves**

118   Figure 4 shows the complete set of dispersion curves obtained in this study, after quality

119   control was performed. A total of 105 curves was measured, at periods ranging from 16 to 333 s

120   depending on path length and data quality; generally speaking, longer periods were measurable

121   on longer paths and shorter periods on shorter paths. We do not, however, have high confidence

122   in the longest periods, which are sparsely sampled, and therefore only periods up to 200 s are

123   shown in the figure. The curves all show a velocity that increases with period, as would be

124   expected given that longer periods sample greater depths. In the 40-200 second range, which is

125   most sensitive to lithospheric depths, the path-to-path variation covers a range of ≈ 400 m/s. This

126   variation is systematic with path location, with eastern paths being faster and western paths being

127   slower.

128   The spatial variation of phase velocities is more apparent in map view (Figure 5). At

129   relatively short periods (up to 50 s), low velocities are clearly restricted to the Cordillera, while

130   moderate to high velocities occur further east. At longer periods, more paths sampling both areas

131   are present; though there is still a trend of velocity increasing eastward, there are also a number

132   of crossing paths with distinctly different velocities. The true distribution of phase velocity

133   cannot be determined without tomographic inversion.

134   **3.2 Tomographic resolution and maps**

135   Under the assumption that two-station measurements represent averages along great-

136   circle paths, the recovery of 2-D dispersion maps is a linear inverse problem (Montagner 1986).

137   We solved for a grid of phase velocities spaced $1°$ apart in latitude and $2°$ in longitude, with

138   smoothing and damping constraints; the correct regularization level was determined by the L-

139   curve method (Parker 1994). Each period was solved for independently, yielding maps from 30

140   to 160 s at 10 s intervals.

141   We performed lateral resolution tests using the path coverage for three relevant periods

142   (50, 80, and 120 s; Figure 6). For an input model consisting of $4°$ (latitude) by $8°$ (longitude)

143   blocks of alternating positive and negative 0.4 km/s velocity perturbations, we calculated average

144   velocities for each path, added errors comparable to the scatter in the real data, and inverted

145   using the same parameters as for the real data. The size of the blocks used reflects the feature

146   size we are able to recover: ~ 450 km over most of the map area. The results indicate that the

147   boundaries of features of this scale can be recovered accurately (within ~ 100 km) west of ~

148   100°W and south of 70°N, albeit with significant underestimation of anomaly magnitudes, and

149   that resolution is best at moderate periods (60-100 s). Given that our regularization is smoothing-

150   based, larger-scale features than this will be recovered more accurately.

151   Final dispersion maps for the real data set are shown in Figure 7. The Cordillera is low-

152   velocity at all periods, with the edge of the low-velocity region corresponding closely to the

153   Cordillera/craton boundary at periods of 80 s and less. At longer periods, the edge is more

154   complex, with low velocities extending some distance into the craton north of 52°N. The

155   Cordilleran low-velocity region is truncated north of 65°N at all periods; its southern and

156   western boundaries are not imaged by this study.

157   East of the Cordillera, phase velocities are moderate to high (e.g. > 4.2 km/s at 80 s

158   period). At short periods, resolution does not extend much east of 110°W due to a lack of high-

159   frequency measurements along the longer paths in the eastern portion of the model. At periods of

160   80 s and higher, the highest phase velocities detected are at the eastern edge of the resolved

161   region, ca. 95°W. Localized high-velocity zones are detected beneath Great Slave Lake (at the

162   southern tip of the Slave Province) and along the southern edge of the model ca. 110°W.

163   3.3 Depth inversion and cross-sections

164   Rayleigh-wave period is often used as a depth proxy in surface-wave studies due to the

165   increase in depth sensitivity with period; however, as any given velocity measurement is

166   sensitive to a large range of depths, phase velocities must be inverted in order to constrain

167   seismic velocity as a function of depth. The depth sensitivity of Rayleigh-wave phase velocities

168   is model-dependent, requiring a nonlinear depth inversion.

169   We inverted for depth using the widely-used Computer Programs in Seismology package

170   (Herrmann 2013), using phase velocities extracted from the tomographic maps, at periods

171   ranging from 30 to 160 s at 10 s intervals, the range of periods at which we are most confident in

172   the map coverage. We used a starting model simplified from IASP91 (Kennett and Engdahl

173   1991), with fixed layer thicknesses ranging from 15 to 25 km (including a two-layer crust 35 km

174   thick; see Figure 8), and inverted for S velocity over 5 iterations, with P velocity determined

175   using a Poisson's ratio fixed to match the base model values (which range from 0.20 to 0.29),

176   and density calculated from the P velocity. Smoothing regularization was used to compensate for

177   the non-uniqueness of the inversion and to encourage model simplicity.

178   Synthetic tests of the depth inversion are shown in Figure 8. For each test, an input model

179   containing a large (300 m/s) perturbation from IASP91 was used to generate a dispersion curve,

180   using the same period sampling as the real data. The dispersion curve was then inverted (using

181   the same inversion parameters as the real data) to test recovery of the input. We find that low-

182   velocity features are significantly smeared, but not missed; high-velocity features at relatively

183   shallow depth (above 150 km) are recovered, but produce a weaker low-velocity artefact below,

184   while deep high-velocity anomalies are poorly recovered. These effects are a consequence of the

185   non-linearity of surface-wave depth inversion, and will be less severe for weaker velocity

186   perturbations.

187   To generate cross-sections, we spatially interpolated phase velocities at 50 km intervals

188   from all dispersion maps, along five great-circle trajectories (Figure 9). The resulting dispersion

189   curves were then inverted to form 1-D S velocity models, as described above; the 1-D models

190   were concatenated to form a 2-D model along each transect. An example of this procedure is

191   given in Figure 10, in which the series of dispersion curves is presented as a pseudosection (top

192   panel) and inverted to obtain a velocity model (presented in both relative and absolute velocity

193   terms). It is worth noting that the lack of short-period sampling in some areas affects the

194   inversion; northeast of the 1100 km point along the transect, the shortest periods are not available

195   (due to longer two-station paths in this region) and the shallowest part of the model is not well

196   constrained. As a consequence, northeast of 1100 km, velocity anomalies appear weaker due to

197   upward smearing of structure into the poorly-sampled shallow depth range. We therefore expect

198   that mantle velocity anomalies will be somewhat underestimated in the more poorly-

199   instrumented portions of the study area; this effect is most pronounced on the AA-AA' section

200   shown in Figure 10.

201   Transects were selected to examine the Cordillera-craton transition along three near

202   boundary-perpendicular sections (AA-AA', BB-BB' and CC-CC'), along-strike variation within

203   the Cordilleran low-velocity anomaly (DD-DD'), and a high-velocity feature detected near Great

204   Slave Lake in the dispersion maps (EE-EE'). The Cordillera is underlain by a low-velocity

205   feature (marked as A on the cross-sections; Figures 11 and 12) averaging 2-4% below IASP91

206   velocity down to ~250 km depth. The adjacent cratonic lithosphere (B) is underlain by high-

207   velocity lithosphere (5-8% above IASP91) down to 200-250 km depth; the contact between these

208   two anomalies is quite sharp (~ 8-10% over 200-300 lateral km at lithospheric depths), lies

209   directly beneath the crustal contact, and is not vertical. In the northern two cross-sections (CC-

210   CC' and BB-BB'; Figure 11) the contact dips 10-20° to the northeast, with cratonic lithosphere

211   overlying asthenospheric material associated with the Cordillera; by contrast, the southernmost

212   section (AA-AA') shows a dip of ~ 30° in the opposite direction. The depth extent of Cordilleran

213   low velocities varies somewhat along strike, shallowing to ~ 150 km depth beneath southern

214   British Columbia (DD-DD'; Figure 12). Our synthetic tests showed that high velocities at

215   lithospheric depths can produce low-velocity artefacts below; however, the artefacts are

216   significantly weaker than the high-velocity anomalies they underlie, which is not the case where

217   feature A underlies feature B.

218   A localized high-velocity zone (feature C) centred on Great Slave Lake (62°N, 115°W) is

219   visible on longer-period dispersion maps (Figure 7). The two cross-sections that intersect this

220   zone (BB-BB' and EE-EE'; Figures 11 and 12) show a high-velocity feature from ≈ 125-275 km,

221   extending somewhat deeper than is typical for cratonic lithosphere in North America (200-250

222   km; Eaton et al. 2009, Yuan et al. 2014). Even deeper high-velocity features (labelled D) are

223   seen in patches beneath the Cordillera and adjacent areas.

224   **4. Discussion**

225   The Cordilleran mantle velocities we have measured are very low – for instance, at 400

226   km along the BB-BB' transect, the lowest mantle S velocity attained is 4.25 km/s from 155-170

227   km depth. Though a very low value for this depth (IASP91 reaches 4.51 km/s at 160 km), it is

228   consistent with the models of Van der Lee and Frederiksen (2005) and Schaeffer and Lebedev

229   (2014), which have comparable velocities in the same location. Based on the velocity-

230   temperature calculations of Hyndman et al. (2009), this low velocity would correspond to a

231   temperature of ~ 1200-1250°C, which is in accordance with previous studies, implying

232   asthenospheric material underlying a weak, thin lithosphere and raising the possibility of small-

233   scale convection in the Cordilleran upper mantle (Hyndman 2010).

234   Though our measurements of Cordilleran upper-mantle velocity primarily confirm earlier

235   studies, this study provides novel high-resolution constraints on a significant portion of the

236   Cordillera/craton boundary. Our cross-sections show that the mantle boundary is not vertical

237   (Figure 11), with some high-velocity material extending west of the Cordilleran deformation

238   front in the crust. This may help to explain why some models place the boundary west of the

239   front (see e.g. Frederiksen et al. 2001, McLellan et al. 2018).

240   Our most remarkable observation at the Cordillera/craton boundary is the change in dip

241   direction from northeastern (cratonic material overthrusting Cordilleran asthenosphere) on the

242   northern and central cross-sections (CC-CC' and BB-BB', Figure 11) to southwestern

243   (Cordilleran asthenosphere overthrusting cratonic lithosphere) in the southernmost cross-section.

244   This change, which was also detected in a recent North American model (Schaeffer et al. 2017),

245   is seen in previous smaller-scale models. Surface and body-wave models of the southern

246   Cordillera (Bao et al. 2014, Chen et al. 2017) show a steeply westward-dipping or vertical

247   boundary, while models of the northern Cordillera show northeast dips (Cook and Erdmer 2005,

248   Mercier et al. 2009). Models of the United States are not directly comparable due to the

249   influence of the Yellowstone hotspot.

250   A possible explanation for the difference in dip direction at the lithospheric contact may

251   be that the northern two cross-sections sample the Wopmay orogen, while the southernmost

252   cross-section intersects the Taltson. Though both of these orogens were assembled through

253   eastward-dipping subduction (McDonough et al. 2000, Davis et al. 2015), their western edges

254   consist of different terranes (the Hottah for the Wopmay and the Buffalo Head for the Taltson).

255   These terranes have different origins, and it is possible that their associated lithospheres are

256   rheologically distinct. Edge-driven convection has been proposed for the southern Cordillera

257   (Hardebol et al. 2012, Bao et al. 2014); modelling results indicate that mantle viscosity is an

258   important control on the behaviour of the Cordillera/craton transition. It is thus plausible that a

259   rheological difference between cratonic lithosphere of different affinity is affecting the shape of

260   the mantle boundary, but further modelling is needed to support this.

261   Localized high-velocity features at sub-lithospheric depths (feature D; Figures 11 and 12)

262   are found at various points beneath the Cordillera and craton. High velocities at these and greater

263   depths appear in a number of tomographic models (see e.g. van der Lee and Nolet 1997b, Ren et

264   al. 2007) and are commonly attributed to remnants of the subducted Farallon or Kula slab. The

265   lateral resolution of our model does not permit an interpretation of the lateral extent of these

266   features east of the Cordillera (as shown by the longest-period resolution test in Figure 6).

267   However, as our path density is highest within the Cordillera itself, the discontinuous nature of

268   feature D in section DD-DD' is probably a real feature, though ~ 350 km depth is at the limit of

269   resolution for fundamental-mode Rayleigh waves.

270   Gaps in subducting slabs, resulting from processes such as ridge/trench contact, are

271   known in the literature as slab windows. Slab windows have been proposed to exist north and

272   south of the Juan de Fuca slab (Thorkelson and Taylor 1989); the southern window is seen in

273   northern California (Hawley et al., 2016), while an additional slab gap is visible as a low-

274   velocity feature beneath Oregon (Tian and Zhao 2012). We interpret feature D as slab remnants

275   interrupted by a northern Cordilleran slab window, and therefore that the slab is absent from 400-

276   1200 km along the DD-DD' cross-section. This range corresponds well with lateral velocity

277   contrasts at ≈ 50°N and 60°N seen in the teleseismic P velocity model of Mercier et al. (2009).

278   The high-velocity feature labelled C in sections BB-BB' and EE-EE' cannot be attributed

279   to recent subduction. Its depth of 150-250 km beneath the stable craton places it at the base of

280   the lithosphere; its velocity of up to 10% over IASP91 is remarkable, indicating that it is fast

281   even relative to the generally high-velocity surrounding cratonic lithosphere. A high-velocity

282   feature beneath the Slave craton was previously found in this depth range using array analysis of

283   surface waves (Chen et al. 2007), while strongly-layered lithosphere is visible in receiver

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284   functions (Bostock 1998). LITHOPROBE reflection data across the Slave craton's eastern boundary

285   show dipping reflectors attributed to Proterozoic and Archean subduction (Cook et al. 1999)

286   which, when extrapolated eastward, correspond with receiver-function arrivals bracketing the

287   depth of feature C. We therefore propose that feature C preserves these subduction remnants in

288   the lower portion of the Slave lithosphere.

289   **5. Conclusions**

290   Through two-station analysis of surface-wave records, we have mapped Rayleigh phase

291   velocity across western Canada. The most salient feature of our dispersion maps is the sharp

292   contrast between low-velocity Cordilleran upper mantle and high-velocity cratonic lithosphere.

293   By extracting dispersion curves from the tomographic maps and inverting for S velocity

294   structure, we have generated cross-sections examining the Cordillera/craton mantle boundary

295   and surrounding region; these cross-sections show a change in dip of the mantle contact from

296   northeast in the northern Cordillera to southwest in the southern Cordillera, which may reflect

297   differences in rheology between the Wopmay and Taltson shield provinces. Isolated high-

298   velocity features seen beneath the Cordilleran upper mantle may represent Farallon slab

299   fragments, separated by a gap resulting from a Cordilleran slab window. A high-velocity feature

300   at the base of the Slave Province lithosphere correlates with a previously interpreted Proterozoic

301   subduction feature.

302   The main limitation on these results is the lack of resolution east of the Cordillera due to

303   a limited station network. New instruments currently deployed in the Northern Cordillera (Ma

304   and Audet 2017) will allow improvement of these images in the future, as will incorporation of

305   instrumentation in the United States.

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306 **Acknowledgements**

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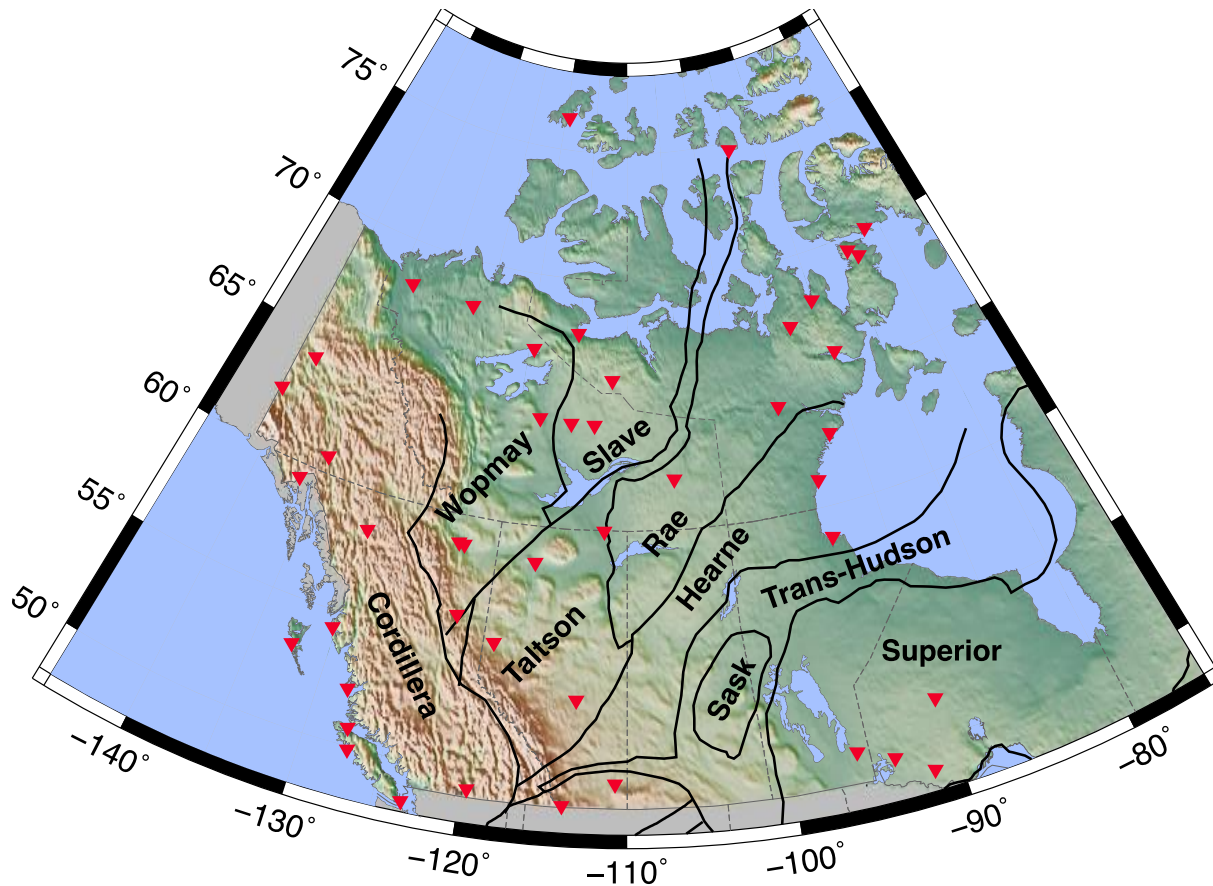314 Artemieva, I.M. 2009. The continental lithosphere: Reconciling thermal, seismic, and petrologic
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437

438   **Figures**

439

440   Figure 1: Seismic stations used in this study (red triangles), overlain on topography. Black lines

441   indicate major tectonic boundaries (simplified from Whitmeyer and Karlstrom, 2007). Map

442   projection: Lambert conic centred at 63°N, 110°W with standard parallels at 53°N and 73°N; the

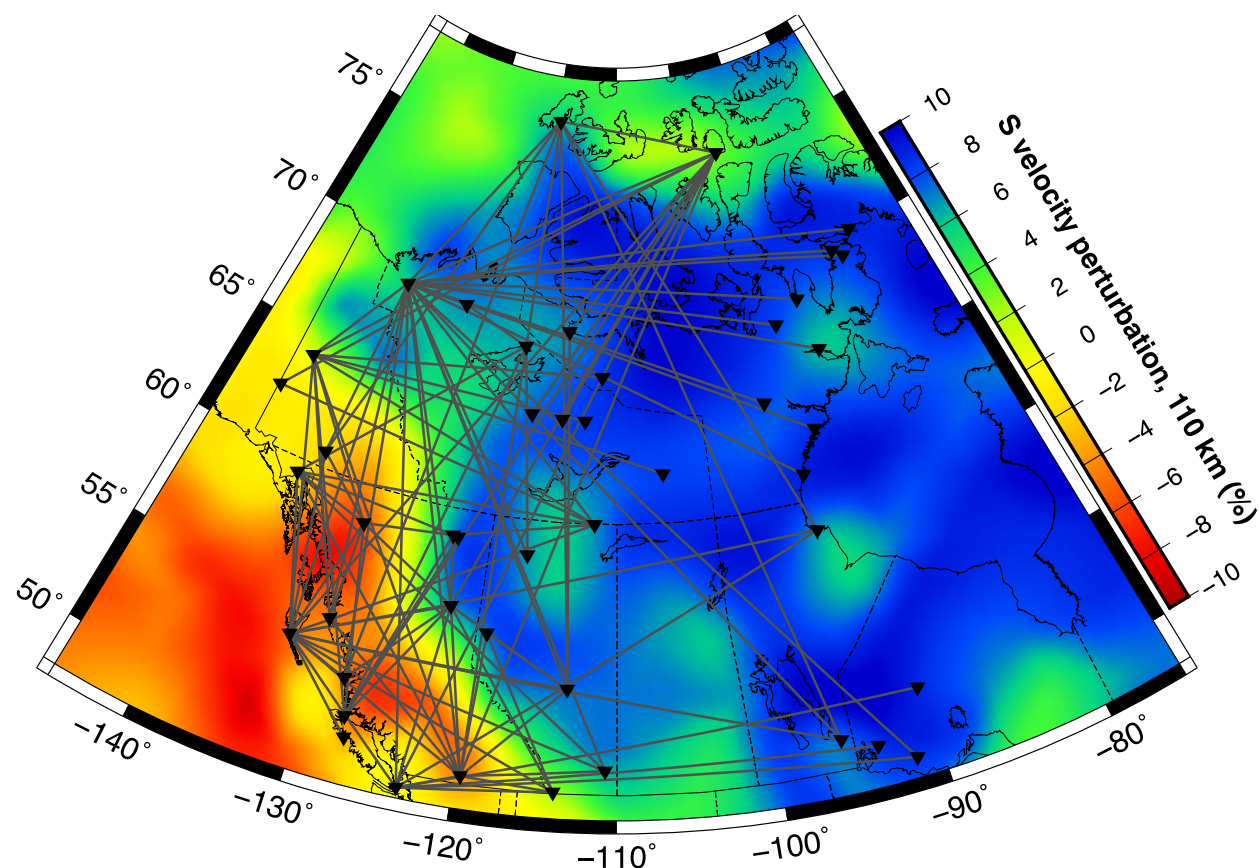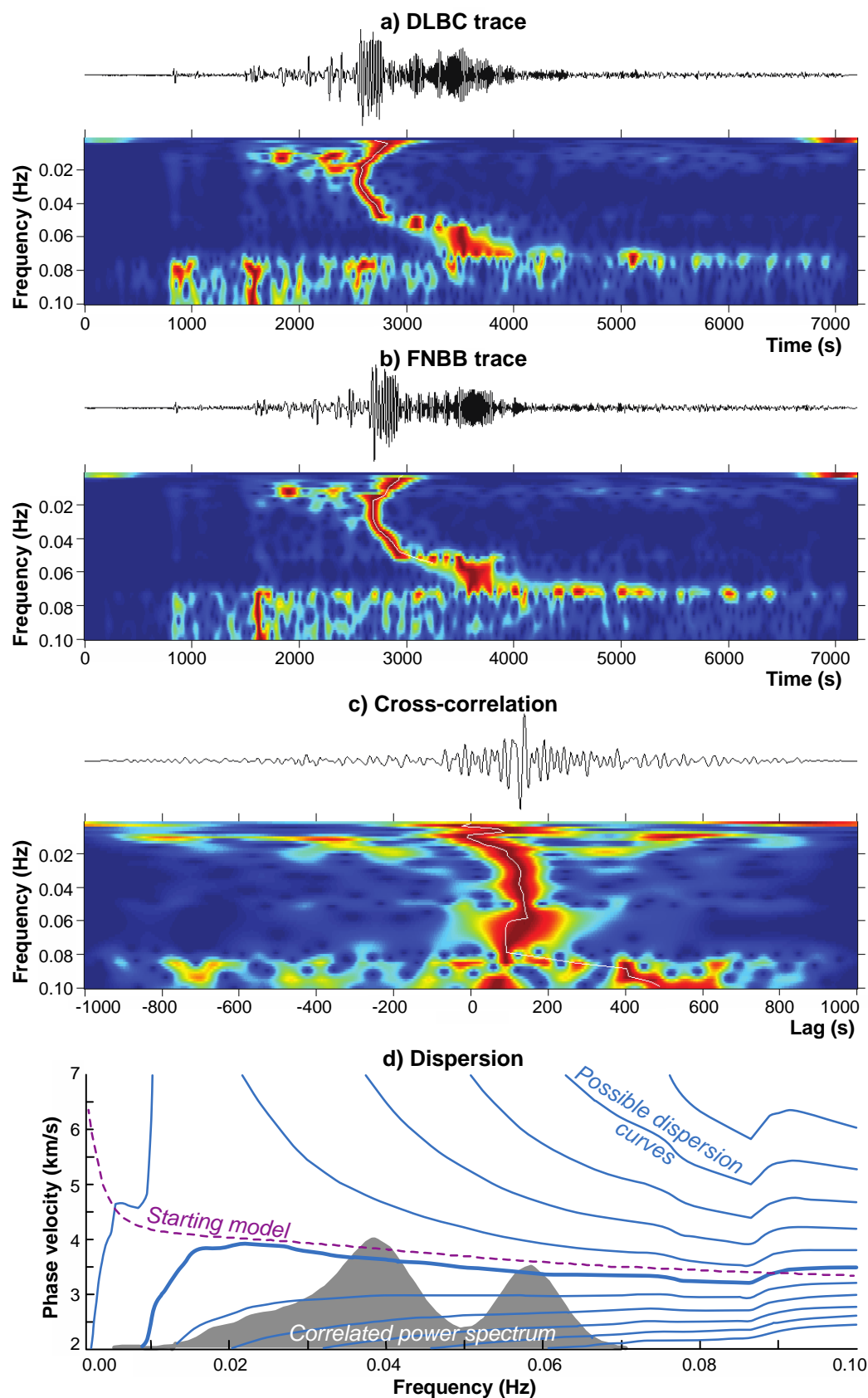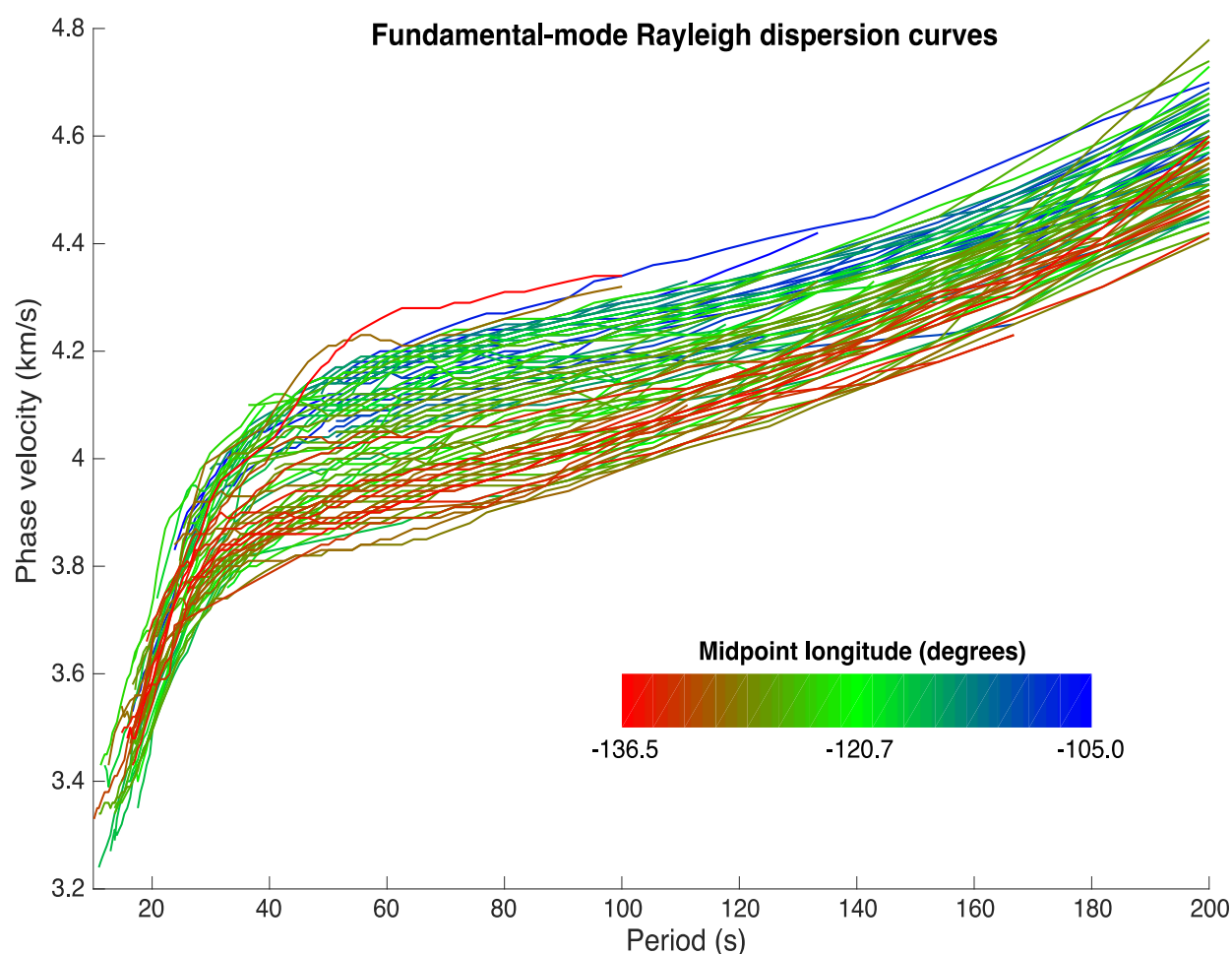443   same projection is used for all other maps in this paper.

444

445   Figure 2: Two-station paths (grey lines) analyzed in this study. Background is seismic *S* velocity

446   at 110 km depth, from the tomographic model of Schaeffer and Lebedev (2014).

a) DLBC trace

b) FNBB trace

c) Cross-correlation

d) Dispersion

447

21

448   Figure 3: Two-station analysis for an event recorded at DLBC (Dees Lake, British Columbia)

449   and FNBB (Fort Nelson, British Columbia). Individual traces (a, b) are shown along with a time-

450   frequency analysis; the thin white line indicates group arrival time. The two traces are cross-

451   correlated (c) and the cross-correlation phase determines possible Rayleigh phase-velocity curves

452   (d, blue lines). Due to the $2n\pi$-radian ambiguity of phase, curves are closely spaced at high

453   frequencies; at lower frequencies, only one is plausible (thick line) and may be followed to

454   higher frequency. Here, the curve is usable from ≈ 0.02 to 0.07 Hz.

455

456   Figure 4: Average phase-velocity dispersion curves for all paths examined in this study. The

457   curves are coloured based on the midpoint longitudes of the corresponding two-station paths.

458

Figure 5: Phase velocities at four representative periods, plotted as shaded lines along the

corresponding two-station great-circle paths. Note that the colour scale differs between panels.
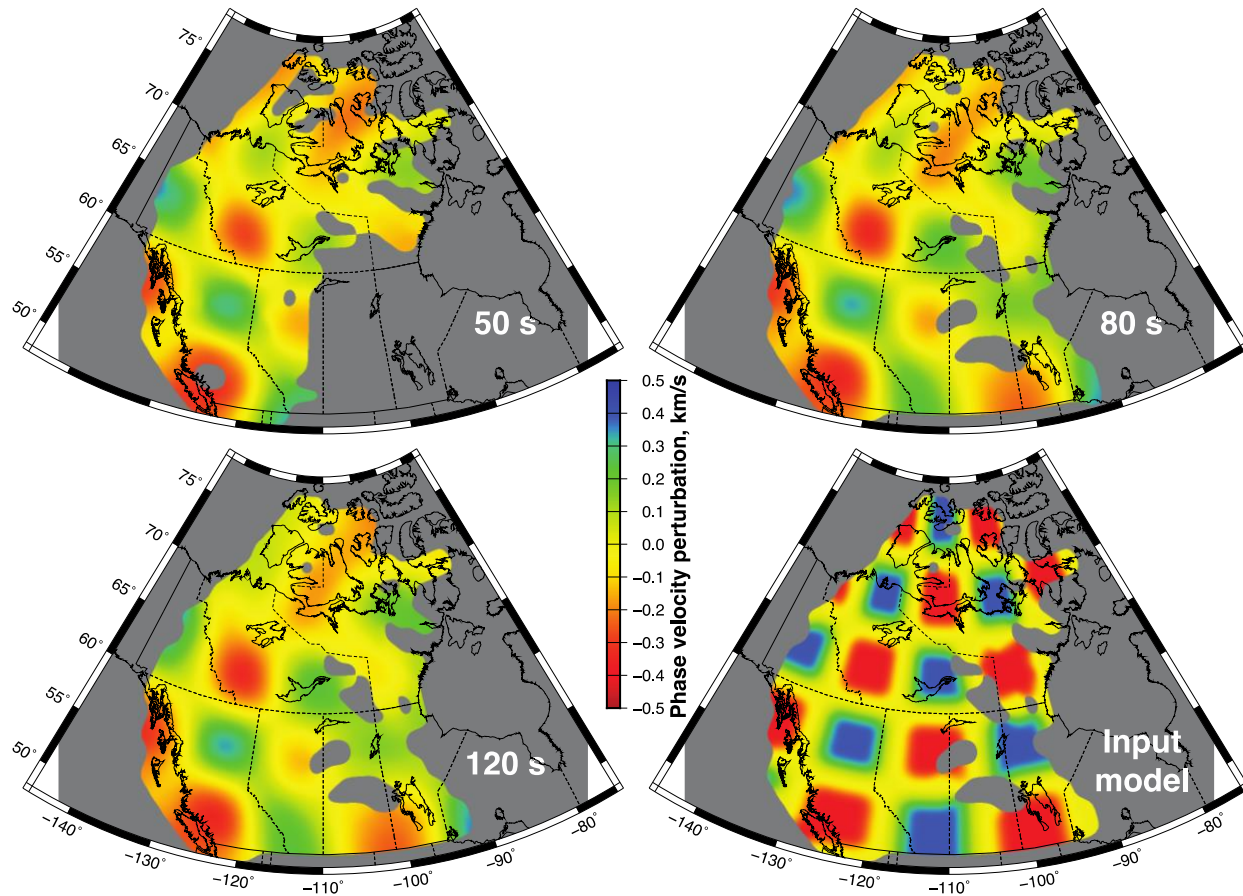
461

Figure 6: Checkerboard resolution tests for phase-velocity tomography, at three representative

periods. Greyed-out model cells were not sampled by rays. The colour scale is the same for all
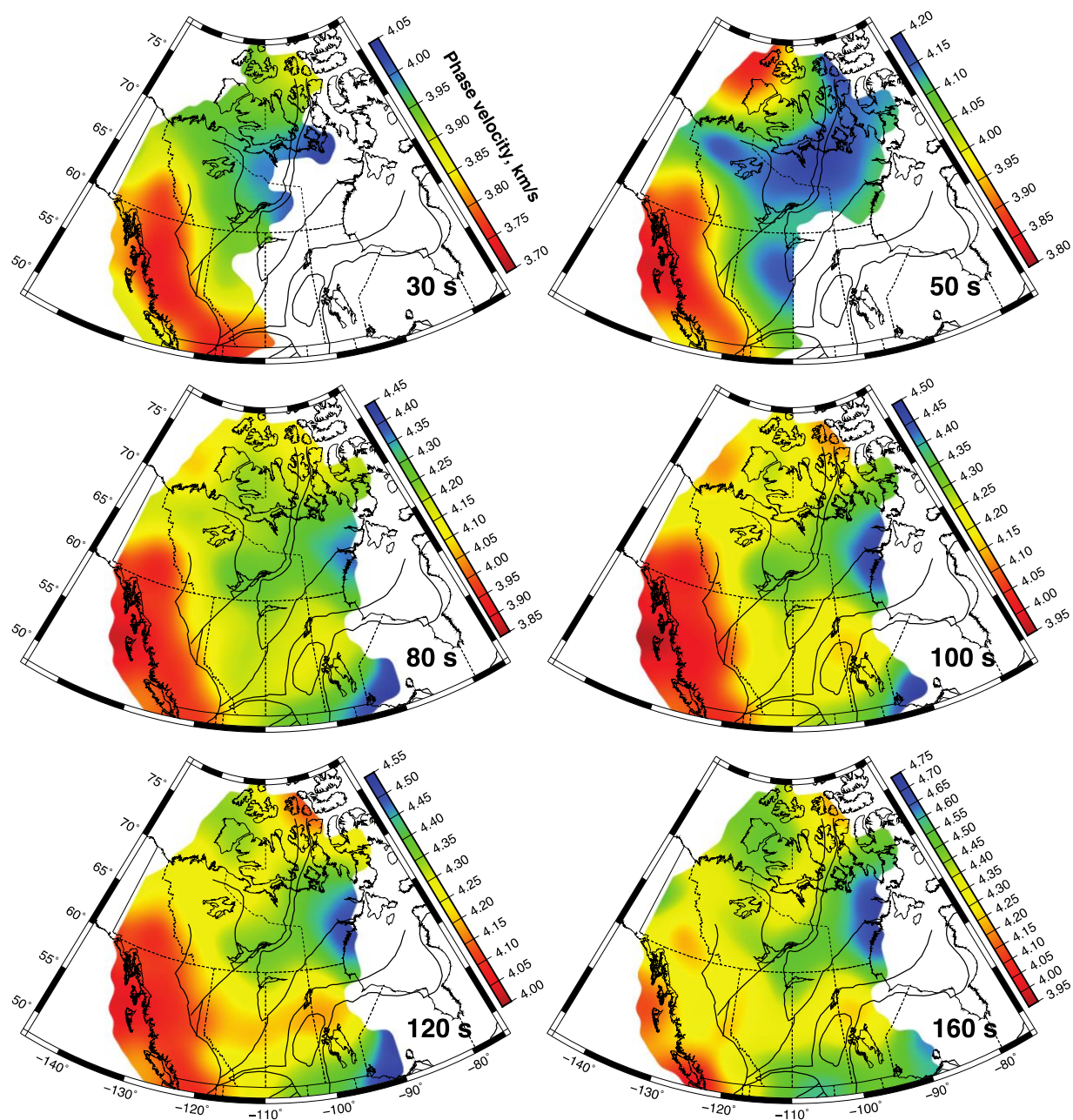
plots.

465

466   Figure 7: Recovered Rayleigh phase-velocity maps at six periods. The colour scale varies

467   between plots. Black lines are tectonic boundaries from Figure 1.

Figure 8: Synthetic tests of depth resolution. The input model (solid grey) contains a perturbation from the background (grey dotted); the recovered model (black dashed) is the result of generating synthetic data from the input model, then inverting the result.
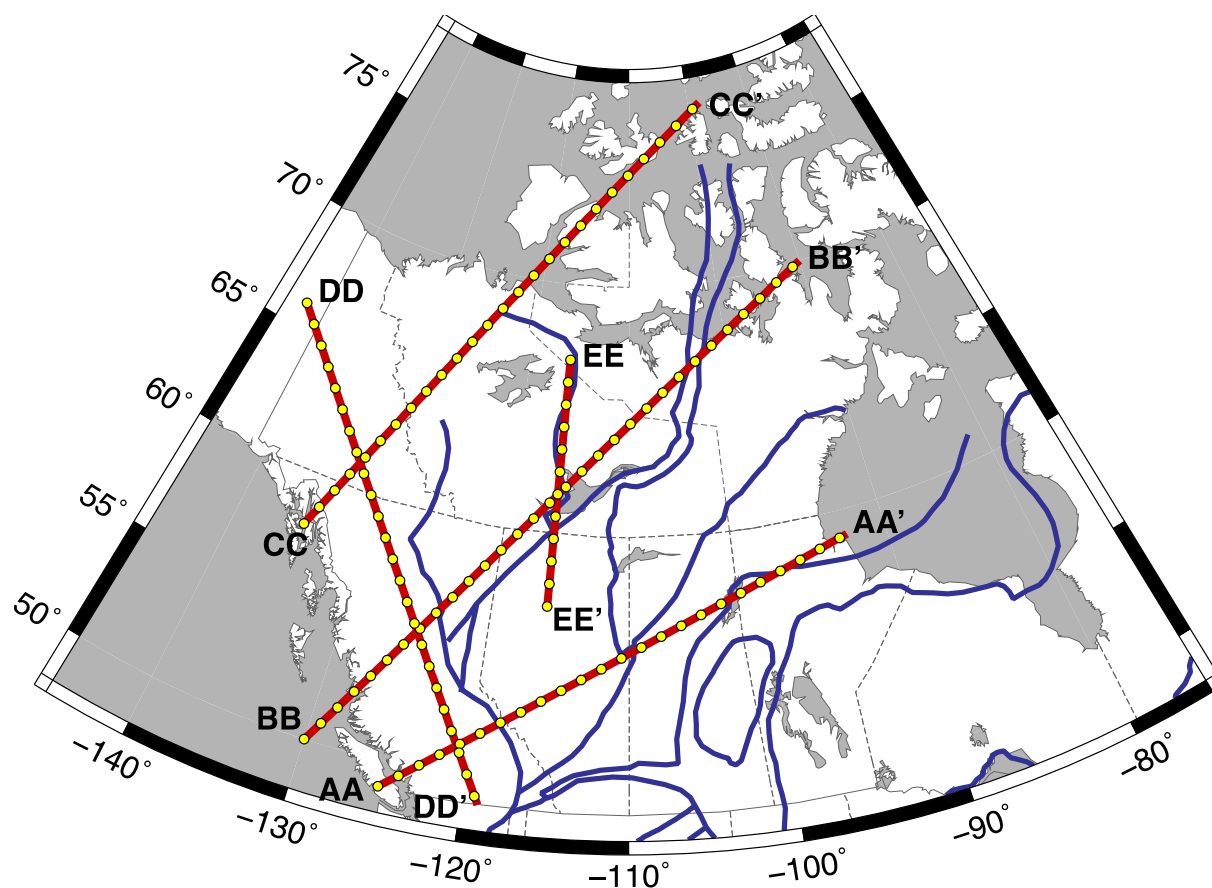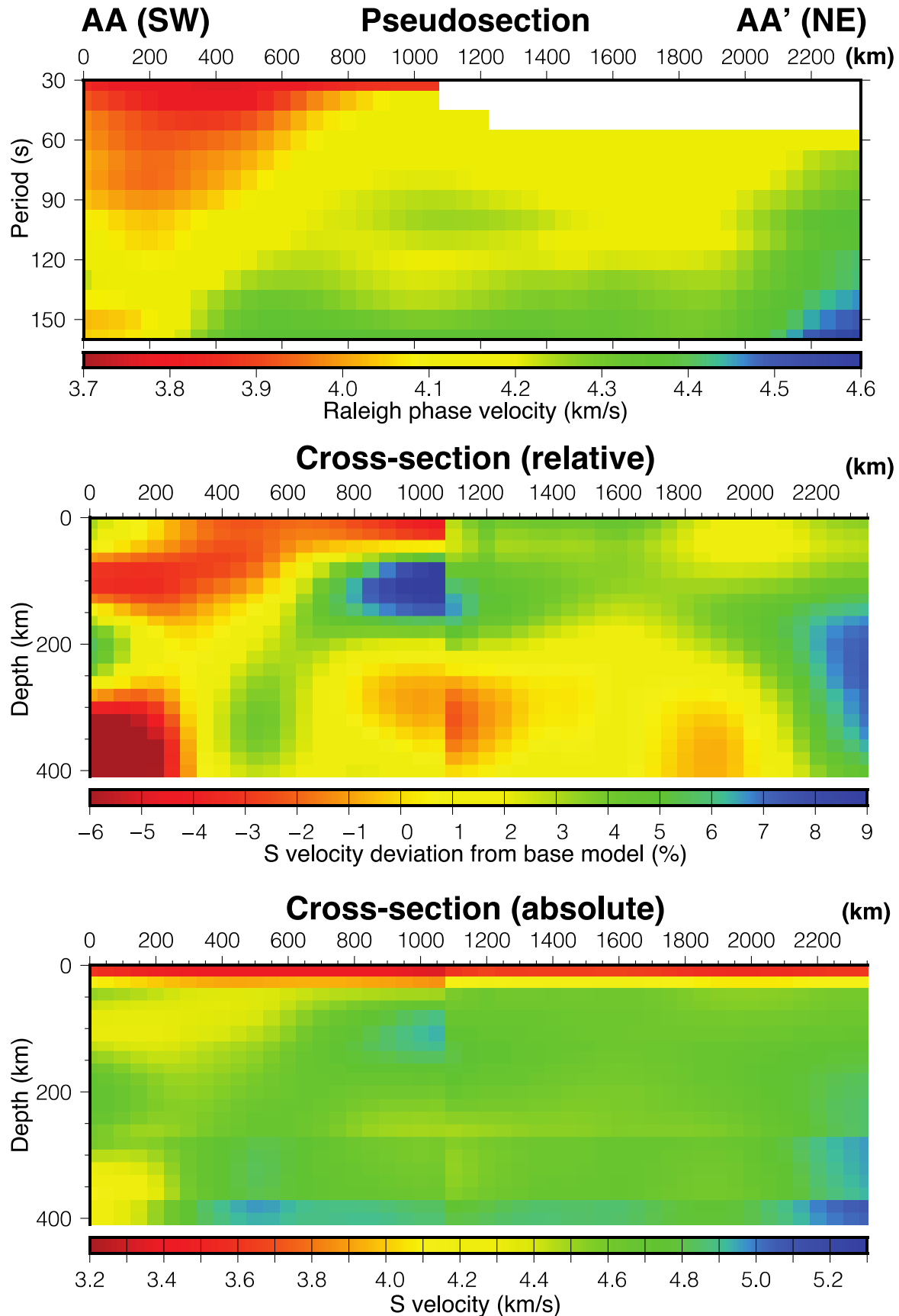
472

473   Figure 9: Locations of transects (red lines) inverted to form cross-sections. Yellow dots mark

474   100 km intervals. The blue lines are tectonic boundaries from Figure 1.

AA (SW)   Pseudosection   AA' (NE)

Raleigh phase velocity (km/s)

Cross-section (relative)

S velocity deviation from base model (%)

Cross-section (absolute)

S velocity (km/s)

475

28

476   Figure 10: Example of the cross-section generation process for transect AA-AA'. A

477   pseudosection (top) is extracted from dispersion maps, then each column is inverted separately to

478   form a 1-D model. Concatenation of these 1-D models generates a cross-section, presented in

479   terms of relative (middle) or absolute (bottom) velocity; the base model is IASP91.
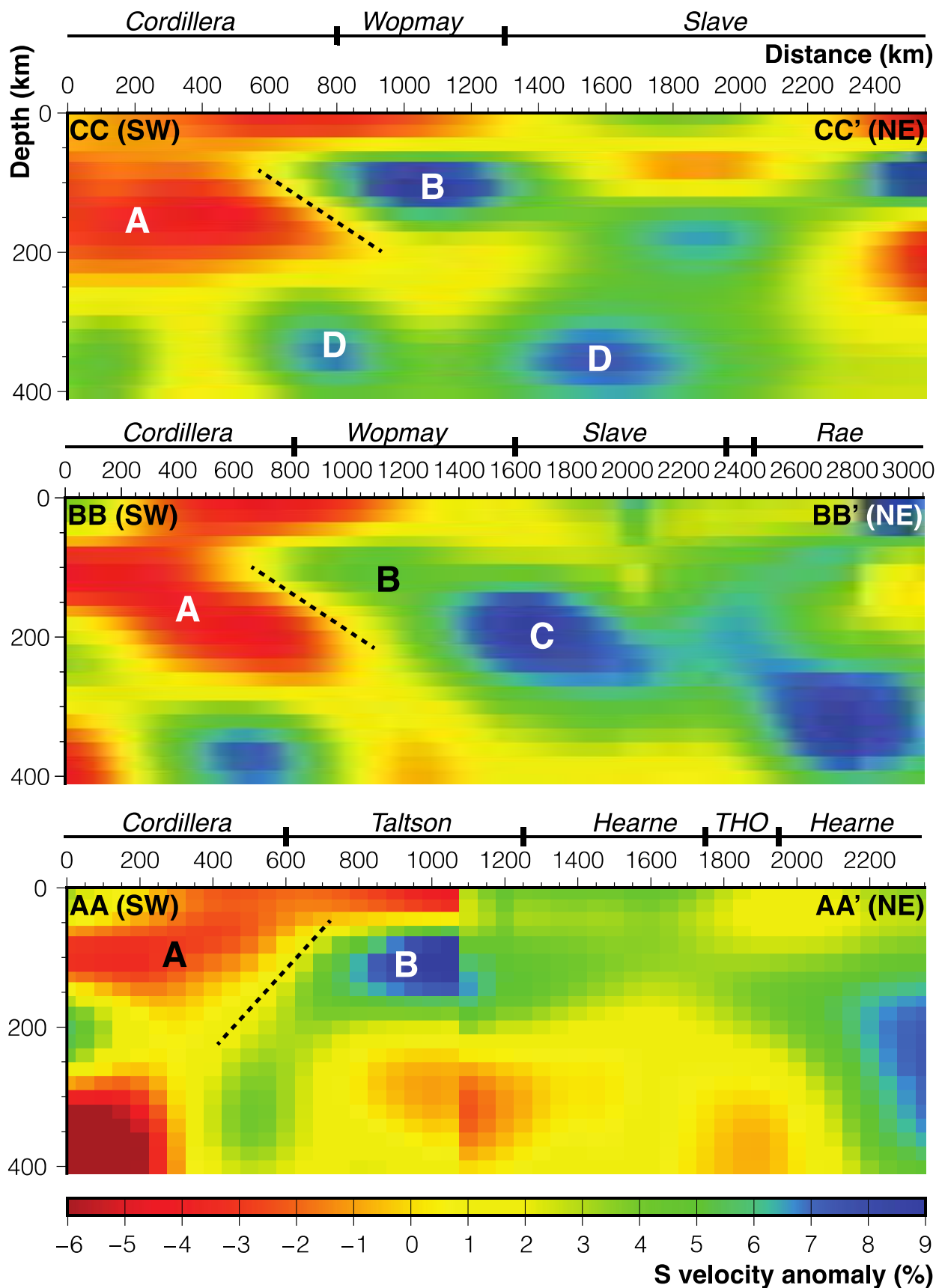
480

481 Figure 11: Three relative-velocity cross-sections perpendicular to the Cordillera-craton

482 boundary, ordered from north to south. Labels at the top of each indicate the tectonic provinces

483 sampled. Letters indicate features discussed in text; dashed line indicates approximate 0%
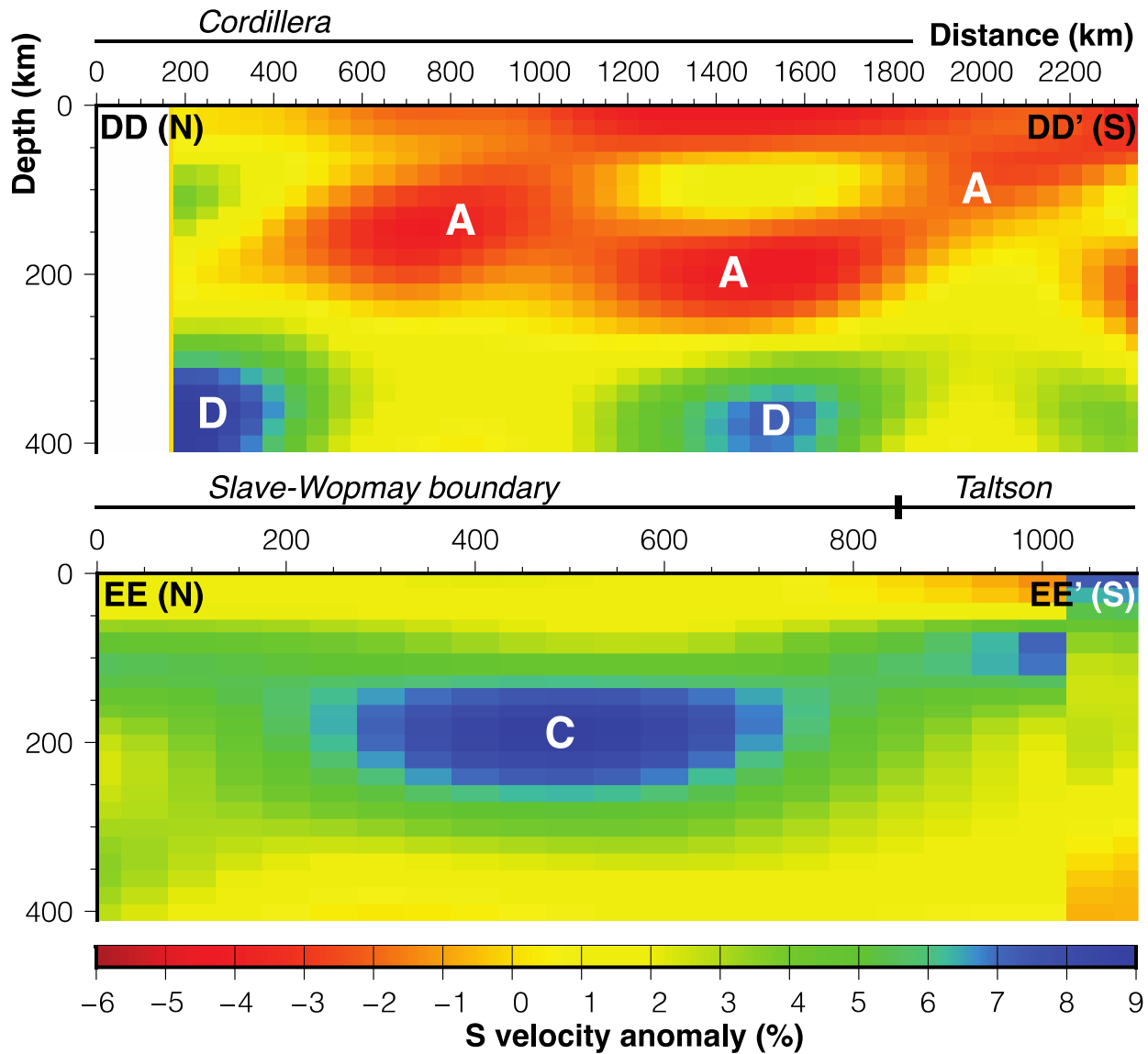
484 contour between features A and B.

485

486 Figure 12: Additional relative-velocity cross-sections along the strike of the Cordillera (top) and

487 across a high-velocity feature (C) near Great Slave Lake.

31